

GEOPHYSICAL METHODS

FUNDAMENTALS, APPLICATIONS, AND CASE STUDIES

O.P. Mishra and D.C. Naskar



Geophysical Methods

This book includes a complete background on geophysical methods of exploration, practices, and case histories for a better understanding of the subject of geophysics and its applicability in diverse fields of exploration. It details both conventional and advanced geophysical techniques, with descriptions of the physics involved in different methodologies. Divided into 16 chapters, the book includes detailed discussions of the theory of individual methods, the operation of specific instruments, the presentation of results, and their interpretation.

FEATURES:

- Discusses potential geophysical methods and applications in mineral exploration
- Reviews natural hazard risk mitigation using geophysical methods
- Covers surface, air, marine, and well logging geophysical applications in natural resource exploration
- Includes electrical, electromagnetic, seismic, and radioactive geophysical methods supported by successful case histories
- Strengthens mathematical and problem-solving skills covering all the geophysical aspects

This book is aimed at graduate and post-graduate students in applied geophysics, exploration geophysics, marine geophysics, engineering, and environmental geophysics.



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The authors dedicated this book to their beloved parents

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(Late Dr. J. Mishra & Late Smt. Girija Devi Mishra)

D.C. Naskar

(Late Shri Sudhanya Naskar & Smt Mongola Naskar)



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Contents

Foreword	xxi
Preface.....	xxiii
Author Biographies	xxix
Acknowledgements	xxxix

Chapter 1 Geophysical Methods: Fundamentals, Applications, and Case Studies 1

1.1	Introduction	1
1.2	Broad Description of Earth Exploration.....	3
1.3	Primary Objectives and Accomplishments	3
1.4	Methodology and Instrumentation	4
1.4.1	Gravity Method	4
1.4.2	Magnetic Method	5
1.4.3	Seismic Method.....	6
1.4.4	Electrical Method.....	6
1.4.5	Radioactive Method	8
1.4.6	Gamma Ray Method	8
1.4.7	Remote Sensing Method	9
1.4.8	Geothermal Method	9
1.4.9	Geochemical Method	9
1.4.10	Excavation, Boring, and Sampling.....	9
1.5	Geological and Geophysical Integrated Interpretation and Modelling Techniques.....	10
1.5.1	Some Successful Case Studies of Geophysical Interpretation.....	11
1.5.2	Regional Basement Modelling	12
1.6	Geotechnical Interpretation of Geophysical Data.....	18
1.7	Geophysical Exploration of Lithium Reserves.....	18
1.8	Integration of Geophysical, Geochemical, and Geological Datasets	19
	References	20

Chapter 2 Environmental and Engineering Geophysics..... 25

2.1	Introduction	25
2.2	Geophysical Methods	28
2.2.1	Geometrical Parameters.....	28
2.2.2	Spatial Distribution of Petrophysical/Geophysical Parameters	28
2.3	Shallow Geophysical Investigation.....	28
2.4	Environmental Problems	28
2.5	Civil Engineering Problems	28
2.6	Environmental Applications.....	29
2.7	Engineering Applications.....	29
2.8	Geophysical Fieldwork	30
2.9	Hydrologic Applications of Geophysical Methods.....	30
2.9.1	Vadose Zone	30
2.9.2	Landfills and Contaminant Transport.....	32
2.9.3	Regional Geology and Coastal Regions.....	33
2.9.4	Geophysical Methods: Hydrogeological Case Studies.....	34

2.10	Determination of Bearing Capacity of Gravel Boulder Deposit from Footing Load Test	37
2.11	Marine Geotechnical and Geophysical Surveys in Complex Geological Formations	38
2.12	Conclusion	39
	References	40
Chapter 3	Potential Field Methods (Gravity and Magnetic)	43
3.1	Gravity Method	44
3.2	Kepler's Laws of Planetary Motion	44
3.3	Basic Principles	46
3.3.1	Units of Gravity	47
3.3.2	Mass of the Earth	48
3.3.3	Gravitational Potential.....	48
3.3.4	Gradients of Gravity	51
3.3.5	Second Derivative	51
3.3.6	Vertical Gradient of Gravity from Top and Bottom of Tower.....	51
3.3.7	Horizontal Gradient of Gravity from Top and Bottom of Tower	51
3.3.8	Shape of the Earth	52
3.3.9	Spheroid and Geoid	54
3.3.10	Variation of g from Place to Place	55
3.4	Measurement of Gravity on Land.....	57
3.4.1	On the Earth's Surface	57
3.4.2	In Boreholes	63
3.5	Measurement of Gravity on Moving Platforms.....	64
3.5.1	Sea Surveys	64
3.5.2	Air Surveys (Accuracies ~1–5 mGal).....	65
3.5.3	Space Measurements	65
3.6	Gravity Surveying	65
3.6.1	Reduction of Gravity Data	67
3.7	Gravity Anomalies	71
3.7.1	Free Air Anomaly	71
3.7.2	Bouguer Anomaly	72
3.7.3	Isostatic Anomaly.....	72
3.7.4	Data Format for Reduction of Gravity Data.....	72
3.8	Isostasy	73
3.8.1	The Airy–Heiskanen Model.....	73
3.8.2	The Pratt–Hayford Model	73
3.8.3	Isostatic Compensation and Vertical Crustal Movements	74
3.9	Rock Densities.....	75
3.10	Interpretation of Gravity Anomalies	77
3.10.1	Regional and Residual Gravity Anomalies	77
3.10.2	Data Enhancement	79
3.10.3	Interpretational Techniques in the Inverse Problem	82
3.10.4	Modelling Gravity Anomalies	85
3.11	Applications of Gravity Method and Case Study	93
3.11.1	Gravity Survey across the Narmada–Son Lineament (NSL), Central India, MP.....	94
3.11.2	Gravity Survey in Lesser Himalayas along the Purnea–Sevoke Transect	95

3.11.3	The Use of Gravity Methods in the Internal Characterization of Landfills: A Case Study	96
3.11.4	Gravity Anomaly of Chromite Deposit Body, Cuba: Case Study	98
3.11.5	Gravity Survey in Devon Area, England: An Outcropping Geological Case Study	99
3.11.6	Possible Triassic Folding below Abu Rassain, the Western Desert, Iraq: A Case Study	100
3.11.7	Application of the Gravity Method in Archaeological Prospection in Jordan: A Case Study	102
3.12	Gravity Data of India for Delineation of Structural Fabric and Crustal Architecture	103
3.13	Interpretation of Gravity Maps and Gravity Signatures of Mineral Deposits: A Review	105
3.13.1	Gravity Maps of Mineral Deposits	105
3.13.2	Gravity Signatures of Mineral Deposits	109
	References	110

Chapter 4 Magnetic Method 115

4.1	Introduction to Magnetic Exploration: Historical Overview	115
4.1.1	Similarities between Gravity and Magnetism	115
4.1.2	Differences between Gravity and Magnetism	116
4.2	Magnetic Monopoles	116
4.2.1	Forces Associated with Magnetic Monopoles	117
4.2.2	Magnetic Dipoles	117
4.2.3	Field Lines for a Magnetic Dipole	118
4.2.4	Units Associated with Magnetic Poles	118
4.3	Magnetization of Materials	119
4.3.1	Magnetic Induction ($B = \mu H$)	119
4.3.2	Magnetic Susceptibility ($k = \frac{I}{H}$)	119
4.4	Mechanisms for Induced Magnetization	119
4.4.1	Diamagnetism	119
4.4.2	Paramagnetism	119
4.4.3	Ferromagnetism	120
4.5	Susceptibilities of Rocks and Minerals	120
4.6	Remanent Magnetization	121
4.7	Magnetic Field Nomenclature	122
4.8	Earth's Magnetic Field	123
4.9	Magnetism and Geology: A Simple Example	124
4.10	Temporal Variations of the Earth's Magnetic Field: Overview	126
4.10.1	Secular Variations of the Earth's Magnetic Field	126
4.10.2	Diurnal Variations of the Earth's Magnetic Field	127
4.10.3	Magnetic Storms	127
4.11	Measuring the Earth's Magnetic Field	127
4.11.1	Mechanical Instruments	127
4.11.2	Magnetometers	128
4.12	Field Procedures	130
4.12.1	Modes of Acquiring Magnetic Observations	130
4.12.2	Magnetic Cleanliness and Interference	131
4.12.3	Strategies for Dealing with Temporal (Drift and Diurnal) Variations	131

4.12.4	Spatially Varying Corrections.....	132
4.12.5	Correcting for Main Field Variations.....	132
4.13	Format for Magnetic Surveying	133
4.14	Data Processing	134
4.14.1	Regional–Residual Separation	134
4.14.2	Derivatives.....	136
4.14.3	Analytic Signal.....	136
4.14.4	Reduced-to-Pole (RTP)	137
4.14.5	Reduced-to-the-Equator (RTE).....	138
4.15	Data Interpretation	138
4.15.1	Qualitative Interpretation	138
4.15.2	Quantitative Interpretation	139
4.16	Presentation of Results (Deliverables).....	139
4.17	Magnetic Anomalies over Simple Shapes	139
4.17.1	Comparison between Gravity and Magnetic Anomalies	139
4.17.2	Magnetic Anomaly over Magnetized Sphere at the Northern Hemisphere.....	141
4.18	Applications of Magnetic Surveying.....	145
4.18.1	Case Study: Central India	146
4.18.2	Case Study: Archaeological Research for Landscape Studies, Kreuttal, Austria.....	146
4.18.3	Case Study, Switzerland: Limitations of Geotechnical-Geophysics Non-Uniqueness.....	147
4.18.4	Case Study: Nickel Sulphide Exploration in the Kambalda District, Western Australia.....	147
4.18.5	Magnetic Anomaly above an Olivine Diabase Dyke, Pishabo Lake, Canada: Case Study	149
4.18.6	Magnetic Method for Archaeological Site Characterization of the Jordan Area, Western Asia: Case Study	150
4.18.7	Geospatial Archaeology, Lothal Revisitation Project, Gujarat, India: Case Study	150
4.19	Geomagnetic Data	151
4.19.1	Geomagnetic Observatories in India.....	151
	References	154

Chapter 5	Direct Current (DC) Resistivity Method.....	157
5.1	Electrical Principles	157
5.1.1	Electric Field and Potential	158
5.1.2	Ohm’s Law	159
5.1.3	Equipotentials.....	161
5.1.4	Atomic Charge	161
5.1.5	Types of Electrical Conduction	161
5.1.6	Potentials in Homogeneous Media.....	162
5.1.7	Single-Current Electrode at Depth.....	163
5.2	Archie’s Law.....	164
5.3	Rock Resistivities	165
5.3.1	General Rules of Thumb for Resistivity.....	166
5.4	Current Distribution	167
5.5	Principles of Resistivity Method	169
5.6	Electrode Configurations.....	174

5.6.1	Wenner Array	174
5.6.2	Lee-Partitioned Array	175
5.6.3	Schlumberger Array	175
5.6.4	Pole–Dipole Array	176
5.6.5	Pole–Pole Array	176
5.6.6	Gradient Array	177
5.6.7	Dipole–Dipole Arrays	177
5.7	Effect of Inhomogeneous Ground	180
5.7.1	Distortion of Current Flow at a Plane Interface	180
5.7.2	Distortion of Potential at a Plane Interface	181
5.8	Resistivity Surveying	185
5.9	Comparison of Wenner, Schlumberger, and Dipole–Dipole Measurements	186
5.10	Problem of Defining Probing Depth	187
5.11	Advantages of Using Logarithmic Coordinates	187
5.12	Goelectric Parameters	187
5.13	Vertical Electrical Sounding	189
5.13.1	Types of Electrical Sounding Curves over Horizontally Stratified Media	189
5.13.2	Electrical Sounding over Laterally Inhomogeneous Media	192
5.13.3	Ambiguities of Sounding Curve Interpretation	192
5.14	2-D and 3-D Electric Imaging	194
5.15	2-D and 3-D Electrical Resistivity Tomography	194
5.15.1	Electrical Resistivity Tomography Field Survey Design and Data Acquisition	195
5.15.2	Electrical Resistivity Tomography Inversion	197
5.16	Applications of Resistivity Survey and Case Study	198
5.16.1	In Groundwater Studies	198
5.16.2	Delineation of Shallow Crustal Structure	199
5.16.3	Delineation of Gondwana Sediments below the Deccan Trap	199
5.16.4	Detection of Subsurface Voids	200
5.16.5	Delineation of Coal-Bearing Gondwana Rocks	200
5.16.6	3-D Configuration of Kimberlite Bodies	200
5.16.7	Deep Crustal Studies	201
5.16.8	Geophysical Imaging of Alpine Rock Glaciers: Murtèl Case Study (Tomographic Goelectric Survey)	202
5.16.9	Geophysical Imaging of Alpine Rock Glaciers: Muragl Case Study (Tomographic Goelectric Survey)	202
5.16.10	The Use of Resistivity Method in the Internal Characterization of Landfills, Ilhavo, Portugal: A Case Study	203
5.16.11	Resistivity Method for Archaeological Site Characterization of Jordan, a Country in Western Asia: Case Study	204
5.16.12	Application of Vertical Electrical Sounding (VES) and Electrical Resistivity Tomography (ERT) in Locating Fresh Water in Coastal Terrain at Vizag Steel Plant (VSP), Visakhapatnam, Andhra Pradesh, India: Case Study	204
Chapter 6	Natural Potentials and Currents	213
6.1	Self-Potential (SP) Method: Introduction	213
6.2	Survey Equipment	215

6.3	Measurement of SP/Field Procedure	216
6.3.1	Potential Gradient Method (Dipole/Leap-Frog/Gradient Configuration)	216
6.3.2	Potential Amplitude, or Total Field Method (Fixed-Base) Configuration.....	217
6.4	Interpretation of SP Data	218
6.5	Applications of Self-Potential (SP) Method with Some Case Histories.....	218
6.6	Self-Potential as a Tool to Assess Groundwater Flow in Hydrothermal Systems: A Review of Volcanology Mapping	220
6.6.1	Volcanoes in Japan: Misti Volcano in Peru, Volcano in Mt Fuji, and Volcano in Miyakejima.....	220
6.6.2	Large-Scale Study on Peruvian and Guatemalan Volcanoes.....	221
6.6.3	Small-Scale Study in the Summit Part of Stromboli Volcano	221
6.6.4	Groundwater Flow Pattern at Vulcano (Italy)	222
6.6.5	Inverse Modelling: The Case of Cerro Prieto (Baja California).....	222
6.6.6	Upflow of Thermal Water along a Shear Fault (Colorado).....	222
6.6.7	Self-Potential Anomalies Associated with the Ascent of Acidic Groundwater	223
6.6.8	Transient Self-Potential Anomaly	223
6.7	Self-Potential (SP) Anomaly of Sulphide Ore Body, Sariyer Area, Istanbul, Turkey: A Case Study	224
6.8	Telluric Currents.....	224
	References	227

Chapter 7 Induced Polarization (IP) Method..... 233

7.1	Introduction	233
7.2	Basic Principle (Polarization of the Earth).....	233
7.3	Phase Domain (Measure Phase Delays between Current and Voltage)	237
7.4	The Magnetic Induced Polarization (MIP) Method.....	239
7.5	Chargeability of Various Materials	239
7.6	IP Instruments	240
7.7	IP Field Survey, Processing, and Interpretation of Data	241
7.8	Cross-Borehole Imaging.....	247
7.9	Limitations of IP.....	247
7.10	Measurement Errors	247
7.11	Applications and Case Histories of IP Survey	248
7.11.1	Mapping Soil and Groundwater Contamination: Identification of a Contamination with Cyanide Complexes (Slags from Plating Works), Czechoslovakia.....	249
7.11.2	Groundwater Exploration Work Near the Sauk-Soo River in Crimea and in the Kalinino Region of Armenia.....	250
7.11.3	IP Interpretation in Environmental Investigations, Crescent Beach State Park, Near Portland, Maine	250
7.11.4	Detection of Metal Pipes and Cables in Baima, China.....	250
7.11.5	Laboratory Case History	251
7.11.6	Mineral Exploration High-Productivity Surveys of Massive and Disseminated Sulphide Ore Body Detection.....	253
7.11.7	MIP Survey over Teck's Highland Valley in British Columbia, Canada.....	253

7.11.8	Geophysical Investigation for Base Metal Mineralization in Karoi–Rajpura Area, Bhilwara District, Rajasthan, by the Geological Survey of India (GSI).....	253
7.11.9	Geophysical Investigation for Copper and Gold Mineralization around the Dhani–Basri and Todi ka Bas Areas, Dausa and Alwar Districts, Rajasthan, by the GSI	254
7.11.10	Geophysical Investigation for Base Metal Mineralization in the Extension Area of the Mamandur Polymetals Prospect in Villupuram and Thiruvannamalai Districts of Tamil Nadu, India.....	255
7.11.11	Geophysical Investigation of the Pb–Zn Deposit of Lontzen–Poppelsberg, Belgium	255
7.11.12	Case Study: An Example of an IP Anomaly over the Durkin Shoot Nickel Prospect, Kambalda Dome, Western Australia.....	256
7.11.13	Gold Exploration Using Induced Polarization (IP) and Resistivity Methods in the Meyduk–Latala Area (Kerman Province): Case Study	256
7.11.14	Application of Geophysics for Mineral Exploration in Small-Scale Mining in Fazenda Santana, Brazil: Case Study	256
	References	258

Chapter 8	Electromagnetic (EM) Methods.....	262
8.1	Introduction	262
8.2	Principles of EM Surveying	263
8.2.1	Faraday’s Law.....	263
8.2.2	Ampere’s Law	264
8.3	Electromagnetic Induction	264
8.4	EM Induction Surveying	267
8.4.1	Horizontal Loop EM Method.....	267
8.4.2	Ground Conductivity Meters (Measurements at Low Induction Number)	269
8.5	EM Methods Using a Plane Wave Source of Excitation	273
8.5.1	The Very-Low-Frequency (VLF) Method	273
8.5.2	VLF Resistivity and Radio Magnetotelluric (RMT, CSRMT) Method	276
8.6	Transient Electromagnetic (TEM/TDEM) Techniques	279
8.6.1	Mineral Exploration	282
8.6.2	Hydrocarbon Exploration	284
8.6.3	Study of a Saltwater Intrusion with Time Domain Electromagnetic (TDEM) Data.....	285
8.6.4	Geophysical Imaging of Alpine Rock Glaciers: Murtèl Case Study.....	286
8.7	Ground-Penetrating or Probing Radar (GPR)	287
8.7.1	Electric Permittivity and Conductivity	287
8.7.2	Electromagnetic Wave Propagation	289
8.7.3	Reflection and Refraction of Plane Waves	291
8.7.4	Scattering and Diffraction.....	291
8.7.5	Horizontal and Vertical Resolution.....	292
8.7.6	Ray Geometry and Travel Times	293
8.7.7	Amplitudes/Attenuation	294

8.7.8	Estimation of Exploration Depth.....	294
8.7.9	Technical Aspects of GPR	295
8.7.10	Data Interpretation	298
8.7.11	Range of Applications of GPR	298
8.8	Deep Probing through Ground-Penetrating Radar (GPR) by Geological Survey of India (GSI): Case Study.....	300
8.9	Delineation of Lead (Pb) Contamination from GPR Data	301
8.10	Development of Handheld Step-Frequency GPR Using Fieldfox Network Analyzer and Its Applications	302
8.11	Geophysical Imaging of Alpine Rock Glaciers by Georadar Survey: Murtèl Case Study	303
8.12	Geophysical Imaging of Alpine Rock Glaciers by Cross-Hole Radar Survey: Muragl Case Study	304
8.13	Archaeological Geophysics	304
8.14	Geotechnical Studies	309
	References	311
Chapter 9	Magnetotelluric (MT) Method.....	320
9.1	Source of Signal	320
9.1.1	Theoretical Considerations.....	321
9.2	Behaviour of Electromagnetic Fields	322
9.2.1	Assumptions of Magnetotellurics.....	324
9.3	Electromagnetic Induction in a Homogeneous Half-Space.....	324
9.3.1	Skin Depth.....	327
9.3.2	Anisotropic Half-Space	327
9.4	Impedance Tensor	327
9.4.1	1-D Earth.....	328
9.4.2	2-D Earth.....	328
9.4.3	3-D Earth.....	329
9.5	Rotation of Impedance Tensor	330
9.6	MT Strike Analysis	331
9.6.1	Skew	331
9.6.2	Tipper	331
9.6.3	Induction Arrows.....	332
9.7	The Effects of Static Shift	333
9.8	3-D Galvanic Distortion and Decomposition of MT Impedance Tensor	333
9.9	Rotation-Invariant Magnetotelluric Impedance Tensors.....	333
9.10	Magnetotelluric Phase Tensors.....	335
9.11	Inversion Routines	336
9.11.1	Overview of Inversion Methods	336
9.11.2	The Model Space Occam and Non-Linear Conjugate Gradient (NLCG) Methods	337
9.11.3	The Data Space Method	338
9.12	Variant of Magnetotellurics.....	338
9.13	Instruments Used in MT.....	339
9.14	MT Field Layout.....	339
9.14.1	Multichannel Configuration	340
9.14.2	Remote Reference Magnetotelluric (RRMT)	340
9.15	Processing and Interpretation of MT Data.....	341

9.16	Applications of MT Method and Case Study	341
9.16.1	Commercial Applications.....	343
9.16.2	Research Applications	345
	References	353
Chapter 10	Seismic Method and Seismic Tomography	358
10.1	Introduction	358
10.2	Seismic Waves.....	358
10.2.1	Stress and Strain.....	358
10.2.2	Types of Seismic Waves	359
10.3	Seismic Wave Velocities	360
10.3.1	Elastic Wave Velocity as a Function of Geological Age and Depth	361
10.3.2	Time Average Equation to Estimate Rock Porosity	361
10.4	Ray Paths in Layered Media.....	362
10.4.1	Reflection and Transmission of Normally Incident Seismic Rays	362
10.4.2	Reflection and Refraction of the Obliquely Incident Rays	363
10.5	Seismic Data Acquisition Systems	365
10.5.1	Seismic Sources.....	365
10.5.2	Transducers	367
10.5.3	Seismograph/Seismometer	367
10.6	Seismic Reflection Method.....	368
10.6.1	Introduction and General Considerations.....	368
10.6.2	Seismic Reflection Field Survey.....	375
10.7	Seismic Refraction Surveying	378
10.7.1	Refraction at a Horizontal Interface.....	379
10.7.2	Refraction at an Inclined Interface.....	381
10.8	Seismic Tomography	383
10.8.1	Theory	383
10.8.2	History.....	384
10.8.3	Process.....	384
10.8.4	Limitations	386
10.9	Applications of Seismological Data in India	387
10.9.1	Seismological Data of the Shillong Plateau and the Assam Valley by the GSI	387
10.9.2	Seismic/Earthquake Studies, Tripura Fold Belt, by the GSI.....	387
10.9.3	Seismic Microzonation of Patna Town by the GSI	388
	References	388
Chapter 11	Radioactivity Method.....	391
11.1	Introduction	391
11.1.1	Radioactivity: A Natural Property of Certain Atoms	391
11.2	Radioactive Decay and Half-Life.....	392
11.2.1	The Nature of Radioactive Emissions.....	393
11.2.2	Types of Radioactivity.....	394
11.2.3	Radioactive Decay Series.....	397
11.2.4	Penetration by α , β , and γ Rays	398
11.3	Units of Radioactivity.....	399

11.4	Radioactive Detector	399
11.4.1	Geiger–Muller (GM) Counter	399
11.4.2	Scintillation Counter	400
11.5	Types of Radioelements and Their Detectability	402
11.6	Applications of Radioactivity Surveys in Geophysical Exploration	402
11.6.1	Geophysical Exploration of Minerals and Metals (Surface)	402
11.6.2	Engineering Geophysics for Deciphering Buried Channel and Flow of Underground Fluids	403
11.6.3	Use of Radioactive Measurements as Earthquake Precursors	405
11.7	Radioactive Dating	407
	References	407
Chapter 12	Geophysical Well Logging or Borehole Logging	409
12.1	Wireline Logging	411
12.1.1	History of Well Logging	411
12.1.2	Electrical Logging	412
12.1.3	Porosity Logs	413
12.1.4	Lithology Logs	417
12.1.5	Miscellaneous Logging	419
12.2	Logging While Drilling	420
12.3	Corrosion Well Logging	421
12.4	Memory Log	421
12.5	Coring	422
12.6	Mud Logging	422
12.7	Downhole and Underground Geophysical Methods for Nickel Sulphide Prospect in Kambalda Area, Western Australia: Case Study	422
	References	423
Chapter 13	Airborne Geophysics	425
13.1	Introduction	425
13.2	Access	426
13.3	Sampling	426
13.4	Interdisciplinary Studies	426
13.5	Precise Position Capabilities and Requirements	427
13.6	Measuring the Earth's Gravity Field	429
13.7	Measuring the Earth's Surface Topography	430
13.8	Very High-Resolution Studies with Airborne Techniques	430
13.8.1	Airborne Electromagnetic Method	431
13.8.2	Airborne Magnetic Method	435
13.8.3	Airborne Radiometric Method	438
	References	441
Chapter 14	Remote Sensing Techniques, Global Positioning Systems (GPS), Geographic Information Systems (GIS)	444
14.1	Remote Sensing Techniques	444
14.1.1	Overview	445
14.1.2	Types of Data Acquisition Techniques	446

14.1.3	Thrust Areas in Applications	449
14.1.4	Data Characteristics	449
14.1.5	Data Processing	449
14.1.6	Hyperspectral vs Multispectral Remote Sensing	450
14.1.7	Applications of Remote Sensing in Geophysical Exploration Survey and Case Study.....	451
14.2	Global Positioning System (GPS).....	455
14.2.1	Introduction	455
14.2.2	Accuracies with GPS and Comparison with Other Techniques.....	455
14.2.3	Applications of GPS	456
14.2.4	GPS in India	456
14.2.5	Applications of GPS Survey in Geophysical Exploration Survey and Case Study.....	458
14.3	Geographic Information Systems (GIS).....	462
14.3.1	Introduction	462
14.3.2	Applications of GIS in Geophysical Exploration Survey.....	464
	References	466

Chapter 15 Marine Geophysics..... 471

15.1	Introduction	471
15.2	Propagation of Seismic Waves	472
15.3	Sources of Sediments	473
15.4	Geophysical Techniques and Instruments to Survey the Oceans.....	474
15.4.1	Side-Scan Sonar Survey	474
15.4.2	Swath Bathymetry Survey.....	475
15.4.3	Subbottom Profiler Survey	477
15.4.4	Multichannel and 3-D Seismic Reflection Surveying	478
15.4.5	Gravity/Magnetic Survey	478
15.4.6	Marine Resistivity Survey.....	481
15.5	Corrections Adopted for Marine Geophysical Datasets.....	482
15.6	Survey Approaches and Schedules.....	483
15.6.1	Exploratory or Research Vessels.....	483
15.6.2	Line Spacing.....	483
15.6.3	Integration of Temporally Different Data	483
15.6.4	Survey Data Processing Bottlenecks.....	483
15.7	Role of Marine Geophysics in the Exploration of Deep-Sea Nodules and in the Blue Economy	484
15.7.1	Opportunities for India.....	485
15.8	Detailed Applications of Marine Geophysical Surveys and Case Histories.....	487
15.8.1	Marine Geology and Coastal Geophysical Surveys for Seabed Mapping and Mineral Resource Evaluation in India.....	487
15.8.2	Baseline Geophysical Data for Hazard Management in Coastal Areas in Relation to Earthquakes and Tsunamis	488
15.8.3	The Outer Thames River (England) Estuary Regional Environmental Characterisation	489
15.8.4	Detection of the Torre del Greco Submerged Volcanic Structure at Naples Bay, Southern Tyrrhenian Sea, Italy.....	489

15.8.5	Glacial and Glacially Related Features on the Continental Margin of North-West Ireland.....	489
15.8.6	Gravity Survey at the Oates Coast of Antarctica during the German Antarctic North Victoria Land Expedition (GANOVEX) VIII.....	490
15.8.7	Marine Gravity Survey of Marine Field: Case Study for Silurian Reef Exploration.....	490
15.8.8	Marine Magnetic Survey of a Submerged Roman Harbour, Caesarea Maritima, Israel.....	491
15.8.9	Unexploded Ordnance Located by Marine Magnetic Survey, Tobruk, Libya.....	491
15.8.10	Buried Volcanic Structures in the Gulf of Naples (Southern Tyrrhenian Sea, Italy) Resulting from High-Resolution Magnetic Survey and Seismic Profiling.....	491
	References	492

Chapter 16 A Comparative Overview of Geophysical Methods with Equation

	Dictionaries, Field Layouts, and Instrumental Setups	495
16.1	Overview of Geophysical Methods	495
16.2	Fundamental Properties of Sources	495
16.3	Some Mathematical Identities	498
16.4	Equation Dictionaries in Introductory Geophysics	498
16.4.1	Gravity.....	498
16.4.2	Magnetic.....	504
16.4.3	Seismology	507
16.4.4	Electric and Electromagnetic	513
16.4.5	Magnetotelluric (Natural Source Low-Frequency EM).....	521
16.4.6	Marine Geophysics.....	526
16.5	Physical Properties of Earth Materials.....	527
16.6	Comparison of the Electrical Methods.....	528
16.7	Multiple Methods and Survey Design.....	528
16.8	General Aspects of Geophysical Fieldwork	529
16.8.1	Choice of Method.....	529
16.8.2	Instrumentation	530
16.9	Instrument at the Field Site.....	530
16.9.1	Specific Operational Precautions	530
16.10	Selection of Survey Parameters.....	530
16.10.1	Traverse or Line Spacing.....	531
16.10.2	Station Spacing.....	531
16.10.3	Orientation of Lines	531
16.11	Field Readings	531
16.11.1	Sensitivity, Precision, and Accuracy	531
16.11.2	Errors.....	532
16.12	Base Stations.....	532
16.12.1	Continuous Recording.....	532
16.12.2	Periodic Repeat Reading	532
16.12.3	Gradiometers	532
16.13	Signal and Noise	533
16.13.1	Definitions and Examples	533
16.14	Data Sheet Format for Common Geophysical Methods.....	533

16.14.1 Wenner Electrical Resistivity Data Sheet	533
16.14.2 Schlumberger Electrical Resistivity Data Sheet.....	534
16.14.3 Collinear Dipole–Dipole Electrical Resistivity Data Sheet (Traversing)	534
16.14.4 Dipole–Dipole Electrical Resistivity Data Sheet (Sounding, Azimuthal Technique)	535
16.14.5 Dipole–Dipole Electrical Resistivity Data Sheet (Sounding, Equatorial Technique)	536
16.14.6 Self-Potential (SP) Data Sheet.....	537
16.14.7 Induced Polarization (IP) Data Sheet.....	537
16.15 Determining the Anomaly.....	538
16.16 List of Selected Glossary, Acronyms, Symbols, and Terms.....	539
References	541
Index	543



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Foreword

Geophysics is a principal tool for geoscientists that helps decipher not only deeper geology but also formational processes and chronological changes in the physical property contrasts between the causative bodies and the surroundings at the different strata of the subsurface.

It is a fact that geophysical methods are very effective in resolving various problems related to the lithospheric, atmospheric, hydrospheric, and cryospheric domains because of advanced observational systems and powerful computational algorithms for geophysical problems for the entire earth system science. However, in geosciences, the principal problem is that geophysical exploration does not demonstrate the existence of materials and structures directly but is rather indicative of proxies, associated with mapping the distribution of estimated physical parameters. The selection of these parameters and the subsequent interpretations of measured data are key points of every geophysical survey. Another serious drawback of the geophysical method is the non-uniqueness of the interpretation. It is therefore the authors who advocate for the deployment of integrated geophysical methods to find a plausible inference from the generated diverse set of geophysical data, since measurements of several geophysical parameters with broader spectra provide ample opportunity for the joint interpretation of the obtained results that lead to correct and useful findings.

This book, titled *Geophysical Methods: Fundamentals, Applications, and Case Studies*, indeed, reflects a seminal attempt of authors that contains a total of 16 chapters covering both theoretical and applied aspects of geophysics, besides several case studies and references. This book deals with all the current geophysical methods, including electrical; magnetic; electromagnetic; gravimetric; geothermal; seismic; magnetotelluric; borehole or well logging; remote sensing; global positioning system (GPS); geographic information system (GIS); airborne, shipborne, or marine geophysics; and radioactive method. A lucid explanation of the physical principles of each geophysical method provided by the authors in each chapter for easy understanding is one of the essences of this book. A step-by-step derivation of the critical equations with explanatory figures is given for geophysical formulations that extend the geophysical interpretation for oil and gas exploration, mineral prospecting, geothermal exploration, groundwater exploration, engineering applications, archaeological interests, glaciology, volcanology, and environmental concerns. This book includes geophysical survey design, data acquisitions, data processing, and data interpretation with a detailed explanation of the survey study so that even beginners can start with the routine job of geophysicists. A good understanding of mathematics and physics, knowledge of geology and computer skills, and hands-on experience with electronic instruments that geophysicists require are vividly highlighted in this book. In the current scenario, geophysicists are working on various aspects that are covered in this book, which also serves as a textbook for academicians, researchers, and students. In a nutshell, this textbook provides a comprehensive account of the principles, practices, and application of geophysical methods for exploring the subsurface using surface, subsurface, marine, airborne, and satellite-based observation systems, by deploying diverse sets of geophysical equipment.

I take this opportunity to congratulate the authors and the publishers, Taylor & Francis Books (CRC Press), for producing such an excellent book which brings together all geophysical methodologies and their applications that suffice several case studies associated with geophysical surveys in different parts of the world, including of the Indian region. I am sure the book provides a plethora of information along with strong fundamentals on geophysical sciences that will play an essential role in bridging gaps in understanding the efficacies of geophysical methods and their wide-scale applicability under diverse geological and geotectonic conditions.

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Preface

Over several years of engagement in geophysical surveys, teaching, research, and imparting training to professionals and students at different institutes and organizations of India and abroad, the authors often realized a need for a comprehensive text with crystal clear fundamentals on methodologies of geophysical surveys and their applications to a diverse set of problems covering domains of mineral exploration and environmental and engineering geophysics, by adopting surface, airborne, and marine geophysical methods of exploration. To the best of our knowledge, there is not yet any textbook available that adequately covers areas of geophysical applications to explore the shallow subsurface, targeting for mineral exploration in different tectonic environs using integrated geophysical methods consisting of surface, airborne, and marine geophysical techniques to resolve the disposition of the target bodies. Many potential clients either appear unaware of how geophysics could possibly be of help to them or they are having no updated view on geophysical techniques. There has been no suitable book to recommend to them that explains what they need and want to know that requires real examples in a lucid style of understanding the text. Our present work is an attempt to address the gap by providing complete conceptual background to students, researchers, and academicians, professionals, and clients. This book focuses on essential geophysical theories, methods, and techniques in the field mineral exploration, environmental, and engineering geophysics that can contribute to the detection and delineation of subsurface mineral sources with due care of the environmental protection. The idea for this book originated in 1995 while we were working with the Geological Survey of India, Ministry of Mines, Government of India, where we were extensively involved in geophysical field surveys using different potential and non-potential geophysical methods in a very complex and intricate geotectonic settings. The writing of the present book, however, came to fruition after we joined the National Centre for Seismology, Ministry of Earth Sciences (MoES), Government of India, upon the receipt of consistent motivation and inspiration from colleagues and mentors working with different institutes in India and abroad and the encouragement of the present secretary of MoES, Dr M. Ravichandran, to write a textbook for young students, researchers, academicians, and practicing professionals. This is because our teaching to beginners newly recruited to join GSI and exploring other sectors provided us a sound understanding to acquaint them with a solid knowledge base related to the development of new frontiers of geophysics, including instruments, improved data processing, and an idea on geophysical interpretation into geology using software to make them aware about physical processes, which is still continuing at a seemingly ever-faster rate. Much of this has also been fuelled by the availability of evermore powerful computers and associated technology. It has been observed that students who have only very basic mathematical skills with a physics background find difficulty in understanding even the introductory texts in geophysics, owing to the perceived opacity of texts and the daunting display of apparently complex mathematics; consequently, undergraduate and junior master's students treat the interesting subject of geophysics as very hard and boring. This is the reason that we attempt here in this book to remove geophysical phobia from their minds, by adding several illustrations and case studies. The aim of this book is to provide a basic introduction crisply, keeping the mathematics and theoretical physics to a minimal and emphasizing their applications. Considerable effort has been made in compiling a representative set of case histories that demonstrate clearly the issues being discussed.

This book pays attention to a great deal of new material or topics not previously discussed in detail, for example, geophysical survey design and line optimization techniques, image processing of potential field data, recent developments in high-resolution seismic reflection profiling, electrical resistivity subsurface imaging (tomography), electromagnetic along with ground-penetrating radar, magnetotelluric, and induced polarization, among many other subjects. By retaining much of the basic theory and principles from published sources, the scope of materials has been expanded

considerably in this book to reflect the changes and developments in the subject. Consequently, there is much new published material/case histories that have been included, along with geophysical survey methods.

We witnessed development of new branches of the geophysical sciences, such as atmospheric geophysics, forensic geophysics, and agro- and bio-geophysics, as well as techniques to mature, particularly in environmental geophysics and applications to contaminated land, for instance. There has been a move away from just mapping to more monitoring and time-lapse surveys. There has also been a greater blurring of the boundaries between industrial sectors. Hydrocarbon exploration and analytical techniques are now being used in ultra-high-resolution engineering investigations. Electromagnetic methods have ventured offshore, and now it has become a well-established method for hydrocarbon exploration.

It is inevitable with a book of this kind that brand names, instrument types, and specific manufacturers are named. References to such information do not constitute an endorsement of any product, and no preference is implied; nor should any inference be drawn over any omissions. In books of this type, the material covered tends to be flavoured by the interests and experiences of the author, and I am sure that this one is no exception. We hope that all materials presented in this book prove to be beneficial to readers, users, professionals, and students. We would be more than happy to receive any case histories that they feel are of particular significance, which should be for possible inclusion at a later edition. Also, any comments or corrections that readers wish to share us that would be gratefully received for inclusion in future edition.

We attempted to work methodically towards the creation of this book, which contains a complete background on geophysical methods of exploration, practices, and case histories for better understanding of the subject of geophysics and its applicability in diverse fields of exploration. We hope that the information on the pages demonstrate the range of applications of recent trend in geophysics, now very well established, and some of these are very much in the early stages of implementation. It is also hoped that the book may build a strong foundation on which the reader can take the subject further for higher studies, by following the case histories presented for different geophysical methods for dealing with different problems. The references cited, by no means exhaustive, include the aid of addresses to recent websites that have been included to provide pointers to more detailed discussions.

The book is divided into a total of **16 chapters** containing details on various aspects of the geophysical methods and their applications, and a glimpse on each chapter is provide in the following paragraphs. The early **chapters** of this book are an attempt to make a handbook for undergraduate and second- and third-year master's students whose innovative minds may yield new concepts of geophysical processes for extensive applications of geophysical tools for diverse processes. This book details on both conventional and advanced geophysical techniques, with a description of the physics involved in the different methodologies discussed under each chapter, for its better use, which will certainly serve as a reference and text material on geophysical exploration techniques for students and can help in the furtherance of geophysical research in the field of Earth sciences.

Chapter 1 lucidly narrates the descriptions of different geophysical and/or Earth exploration methodologies and their related instrumentations, objectives, applicability, and accomplishments. Then we start with the general description of environmental and engineering geophysics in **Chapter 2**, showing how their interrelationship does exist, with a sufficient highlight on applications of geophysical methods.

Potential field methods are described in **Chapter 3** and **Chapter 4**. **Chapter 3** discusses the gravity method, which involves tiny differences in the gravitational force. Since, according to Newton's law, gravitational force depends on the masses of the bodies, it allows us to differentiate among underground bodies and structures if density contrasts exist between the causative bodies and the host formations. It describes Newton's law of gravitation, gravity meters, gravity field of the Earth, and data reduction, rock densities, gravity effect of selected bodies, gravity data processing with interpretation of results, and applications of the gravity method in the national and international levels.

Chapter 4 includes the basic task of magnetic methods in prospection geophysics, which is to differentiate subsurface according to its magnetic properties. We emphasize the original use of magnetometry in the field of iron ore prospective and demonstrate how magnetometry is being used as a potential method along with gravimetry; hence, they are often measured and interpreted together, which is also discussed in **Chapter 4**. It includes basic principles, like the magnetic field of the Earth, including the main field; the external magnetic field, which covers solar diurnal variations, lunar variations, and magnetic storms; and the spatial variations of the main field, and the magnetism of rocks and minerals, mentioning diamagnetics, paramagnetics, ferromagnetics, anti-ferromagnetics, and ferrimagnetics, and the field instruments for magnetic measurements describing proton precession magnetometer, magnetic surveying with applications on magnetic effect of a sphere and dykes, processing and interpretation techniques, and case histories. All the references are also given for easy understanding of these methods and applicability.

Chapter 5 discusses geoelectrical methods, covering a large group of various methods that differentiate the subsurface according to its electromagnetic properties. There is a wide range of geoelectrical methods, from which we learn only on some simple DC (direct current) resistivity method. We highlight the principal advantage of these methods twofold. First, they are very versatile in dealing with diverse tasks and these can be deployable in the field easily. Second, the simple geoelectrical measuring device is cheaper and easy to build. **Chapter 5** also gives a thorough overview consisting of elementary theory, resistivity of rocks and minerals, resistivity surveying mentioning electrode configurations, vertical electrical sounding, and resistivity profiling. The chapter also highlights the comparison between the Wenner, Schlumberger, and dipole–dipole measurements, the problems of defining probing depth, and the advantages of using logarithmic coordinates. The chapter also contains the most important sample associated with DC resistivity method of geoelectric parameters, types of electrical sounding curves over horizontally stratified media, electrical sounding over laterally inhomogeneous media, limitations of the resistivity method, and analysis of electrical sounding curves. Lastly, applications of resistivity surveys with brief published results with proper referencing have also been provided in the chapter for the readers.

Natural potentials, currents that constitute self-potential (SP), and telluric currents are described in **Chapter 6**, which illustrates the spontaneous polarization or self-potential (SP) methods, where discussion has been made on the measurement of electric potentials developed locally in the Earth by electrochemical activity, electrofiltration activity, or both. The most common use of SP surveys is also discussed in this chapter in the search for ore bodies in contact with solutions of different compositions. The chapter also highlights a broad description on basic concepts, occurrence of self-potentials (mineral potentials, background potentials), mechanism of SP (electrokinetic potential, thermoelectric potential, electrochemical potential, mineral potential), measurement of SP, interpretation of SP data, and then application of SP surveys made in India as case histories, which have publications over important mineralized belts with relevant references for clear understanding and applicability of the method.

Induced polarization (IP) method, systematically described in Chapter 7, is a geophysical imaging technique used to identify the electrical chargeability of subsurface materials, such as ore **bodies**. The polarization effect was originally discovered by Conrad Schlumberger when measuring the resistivity of rock. The survey method is similar to electrical resistivity tomography (ERT) in that an electric current is transmitted into the subsurface through two electrodes and voltage is monitored through two other electrodes. The chapter covers the introduction part, the basic principles, and the origin of IP, describing electrode polarization, membrane polarization, along with constriction within a pore channel, the presence of clay particles or filaments of fibrous minerals, and IP measurement with time domain and frequency domain [e.g., frequency effect, metal factor, relation between time domain (TD) and frequency domain (FD) IP, phase domain, spectral induced polarization (SIP)] measurements. The process of magnetic induced polarization (MIP) method is also discussed, along with the schematic diagram in Chapter 7 in which the benefits from MIP measurement over electric induced polarization (EIP) are also described. The applications of IP survey

in India are also illustrated in the chapter, with published results supported by references cited for it at the end of the chapter.

Electromagnetic (EM) methods do not require direct contact with the ground, as is the case of DC electrical methods, consequently, the fastest coverage of surveyed areas by EM method is easily achieved, which makes EM methods superior over DC electrical methods, which are described in Chapter 8. The chapter covers a large range of EM applications and describes all the necessary steps pertaining to the principles of EM, the depth of penetration of EM wave, the principle of the Slingram method, along with case histories and associated instruments. An attempt is made to undertake discussion in this chapter on the very-low-frequency (VLF) EM method working at various frequency ranges (VLF, 10–30 kHz; LF, 30–300 kHz; and MF, 300–3,000 kHz), and the radio-magnetotelluric (RMT, CSRMT) MT methods (MT, AMT, CSAMT) based on the measurements of the electromagnetic fields of remote radio transmitters. In light of the development of new methodologies of exploration, the technique called ‘ground-penetrating or ground-probing radar’ (GPR) is also discussed in the chapter, covering its principle, data acquisition system, field layouts, resolutions, depth of investigations and penetration, data processing and interpretation of results, limitations, and mode of surveys, with reference to ray geometry along with an exploration depth described with mathematical formulations with suitable case studies, also discussed with the support of the necessary references.

In Chapter 9, the magnetotelluric (MT) method of exploration is discussed in detail, making use of naturally occurring electromagnetic energy propagating into the Earth to determine the electrical resistivity of the subsurface. The chapter describes the fundamental concepts of the MT method based on electromagnetic induction in a homogeneous half-space with Helmholtz equation to estimate the apparent resistivity and phase and anisotropic half-space, applicable for electromagnetic induction in a 1-D Earth, electromagnetic induction in a 2-D Earth, electromagnetic induction in a 3-D Earth, and small-scale and regional resistivity distribution with sub-headings of perturbations of electric and magnetic fields. The chapter also contains some case histories in support of the published references for MT studies.

Seismic methods, with recent advancement of both active and passive sources of seismic tomography, are described in **Chapter 10**. This chapter discusses exploration seismology as the primary method for studying using the concept of vibrational physics with the creation of a source consisting of either natural earthquakes or artificial earthquakes, either through exploding a charge of explosives on the Earth’s surface or through hitting an iron plate with a hammer of sufficient weight, or by vibroseis, which allows sound wave to impinge on the ground and the speed of the wave to get dictated by the media property of strata under the name of seismic waves, used as the most potential method with relatively deeper depth of investigation among all other geophysical methods of exploration. This chapter fully spans several advanced theoretical seismology texts and involves an extensive body of applied mathematics, physics, inverse theory, and empirical knowledge. An attempt has been made to understand that the thrust of theoretical seismology is to develop and improve various techniques for modelling regional and global-scale geophysical processes and to determine the three-dimensional (3-D) structure of the Earth using controlled source (artificial source or man-made) tomography and passive source (natural earthquake) tomography, by estimating both lateral and vertical variations of structural heterogeneities in terms of seismic velocity and several extended parameters from the seismic velocity associated with typical wave propagation phenomena.

It was realized in and around 1945 that there is another important domain of geophysical method of exploration apart from the aforementioned potential and non-potential methods of exploration that depend upon the inherent radioactive behaviour of certain rocks and minerals having significant economic importance, which is described lucidly in Chapter 11. The chapter provides complete information on the basic theory and mathematics of radioactive decay of the methodology and its applications to petroleum and uranium exploration, with a conspicuous demonstration in the chapter about its technological development vis-à-vis sophistication in instrumentation, data processing,

and interpretation of the radioactive results to show how we need these for obtaining resolvable radiograms of the surface Earth's strata.

Chapter 12 describes borehole geophysical measurements through the process of wireline logging under the name geophysical well logging, which allows the investigation and determination of physical properties of rocks at a precisely known depth with varying interval of placing a measuring sensor in the borehole. The fact that downhole logging provides a constant measurement geometry and greater vertical resolution of geological formations is explained, and the possibility of correlation with the drill core exists. The chapter discusses the different types of geophysical methods of logging with their corresponding theory and applications. An illustrative review of borehole logging is also given in the chapter.

With the advent of aviation and space technology, advanced geophysical methods of exploration under the name *airborne geophysical method* have been developed for exploring the Earth, which is discussed illustratively in Chapter 13. The chapter discusses the advantages of the airborne geophysical method over all other geophysical methods in terms of accessibility of hostile terrain, rapid sampling at scales that are optimal for many geophysical problems, and a tremendous potential for interdisciplinary studies at intermediate scales. Different types of airborne geophysical methods, different resolutions, and the penetration of different depths at very low altitudes varying from about 80 to 120 m above the ground, along with the safety and environmental considerations for conducting airborne geophysical surveys, are also highlighted in this chapter.

Chapter 14 presents a geophysical method of exploration that is based on the Earth observation systems of satellite movement around the orbit of the Earth, with various classifications of remote sensing sensor and their different types, data acquisition systems, electromagnetic spectrum, hyperspectral vs multispectral remote sensing, and the thrust area of applications with reference to the Global Positioning System (GPS), the Geographic Information System (GIS), along with requisite software used for processing different dataset, which are all described in the chapter.

Chapter 15 presents a scientific discipline which uses the quantitative observation of physical properties through the use of active and passive sensors to measure the shape and hardness of the seafloor, as well as to define and map sub-seafloor structures, to understand seafloor and sub-seafloor geology. Marine geophysics has also played a vital role in the search for petroleum and other natural resources lying beneath the seabed. This chapter provides a comprehensive review of techniques used to explore the oceans and examines what geophysical observations reveal about the structure and tectonics of the seabed and the Earth's interior.

Chapter 16 highlights the acquisition of data, which is fundamental to geophysical investigations, and 'real' data is only acquired in the field. Without real data, no true practical conclusions can be made about a particular requirement. These notes are largely the result of the authors' over 30 years of experience in exploration geophysics. Some of it is from bitter experience, and it is best to heed the advice. The aim of this chapter is to correlate the basic principles of geophysical methods side by side, starting from the fundamental equations. This study reveals that many analogies exist among these methods, both in their mathematical formulation and, sometimes, in the ways these are used in geophysical applications. The notes are intended to apply in general only with equation dictionaries and to all types of fieldwork with associated instruments, including environmental and engineering applications, which can be different situations from those of other exploration activities. The remarks refer to all geophysical surveying techniques, not to the particular operations of airborne, marine, or underground surveys. Detailed discussions of the theory of individual methods, the operation of specific instruments, the presentation of results, and their interpretation are also included lucidly in tabular format for better understanding.

The subject material has been developed over a number of years, firstly while the authors were entrusted various assignments to accomplish for different regions of India, through their meticulous guidance to the geophysical teams and by involving themselves in translating the best efforts into outputs. Moreover, it has been difficult keeping abreast of all the new ideas, especially with an ever-growing number of scientific publications and the huge resources now available through the

internet. What is exciting is that the changes are still occurring, and we can expect to see yet more novel developments over the next few years. It is our hope that this book will be seen as providing a broad overview of applied mineral exploration, engineering, and environmental geophysical methods, besides the application of geophysical methods in diagnosing the principal causes of uncontrollable natural disastrous events that have tremendous risk potential (e.g. earthquake, landslide, volcanic eruption, tsunamis, avalanches, glacial lake outburst flood [GLOF], etc). The efficacies and sophistication of the various geophysical techniques, along with their limitations, are highlighted. We strongly feel that this book helps in improving the acceptance of geophysical methods and in increasing the awareness of the methods available—then it will have met our objectives. There is no doubt that applied and environmental geophysics have an important role to play, and the potential of geophysical methods of exploration is enormous in the future. All the figures included herein are mostly colour figures, for better legibility. Furthermore, the book will also be available in electronic form in its entirety, and also as e-chapters, all of which will be available for purchase through the website of the publisher.

This book not only offers a valuable resource for geophysical researchers but also demonstrates how geophysics theories and methods can be practically implemented to protect our environment and promote the development of human society and its prosperity. **M/S CRC Press (Taylor & Francis) took immense care and rigorous scrutiny to shape the present book for the best use to various stakeholders.**

We gratefully acknowledge our mentors and senior colleagues with whom we worked and who continuously inspired us to accomplish this task for the benefit of science and the society. We gratefully acknowledge the blessings of our respective parents and the outstanding support we received from our respective families during the preparation and finalization of the book, as without their support, the completion of this book would not have been possible.

O.P. Mishra and D.C. Naskar

Author Biographies

O.P. Mishra is a fellow of the National Academy of Sciences, India (FNASc), for his pioneering work involving applied geophysics, seismology, and disaster risk management, having over 34 years of experience in different fields of Earth sciences and seismology. He obtained his MSc Tech degree from IIT (ISM)–Dhanbad in Applied Geophysics; PhD and DSc from the Geodynamics Research Centre, Ehime University, Japan, in seismology; and post-graduate diploma in management (PGDMA) from the Centre of Education of All India Management Association (AIMA), New Delhi, the apex body of the Indian Institute of Management (s) (IIMs), with specialization in financial management. Dr Mishra worked with IDRC–Canada in a joint-venture project with CMPDIL, Coal India, during 1992–1995, where he worked for detecting and delineating abandoned and unknown coal workings using integrated geophysical methods, comprising both conventional and advanced tools of geophysics; served in the Geological Survey of India for about 20 years (1995–2014), he investigated sub-surface terrains of diverse geological set up in India using different geophysical methods and completed several geophysical projects of repute that yielded several reports and research bulletins to his credit; worked with the SAARC Disaster Management Centre, New Delhi, as the head/director; and worked at GRC Japan as a *foreign* visiting guest professor (November 2005–December 2006). He has the honour of representing SAARC member states as the head and director of SAARC Disaster Management Centre to different UNOs agencies, such as the UNDP, UNESCAP, UNISDR, UNOCHA, and UN-Habitat. Dr Mishra also contributed to IRP, ADPC, ADRC, USAID, and ECO as an invited panellist, contributing significantly in the field of disaster risk management. He is a widely travelled scientist/academician and professional to different countries of the world, having credentials to execute flagship projects funded by the UNISDR (UNDRR) and the World Bank for disaster risk resiliency in South Asia. He is a *life* member of several professional bodies of international and national repute, such as the AGU (USA), SSA (USA), SSJ (Japan), JPGU (Japan), ISCA (India), AOGS (Singapore), IUGG (Germany), MGMI (India), IMS (India), IIPA (India), IGU (India), ISES (India), and AEG (India). He is the recipient of several awards and honours, including a visiting professorship by JSPS (Japan) in the year 2005; the National Mineral Award 2008 for Disaster Risk Management in Applied Geosciences by the Government of India; the A.S. Arya Award 2014 by IIT–Roorkee, Government of India; and the Anni Talwani Gold Medal 2015 by the Indian Geophysical Union. Dr Mishra has supervised and guided more than 50 students and researchers on their higher study during their master’s and doctoral courses, including 10 scholars for PhD degrees from various universities of India. He has more than 150 research publications, including edited books, reports, special issues, and road maps on various themes covering seismology, geophysics, interdisciplinary Earth sciences, and DRR, and in peer-reviewed journals of national and international repute.

Dr D.C. Naskar received his MSc in physics from Calcutta University. He was conferred with doctorate (PhD) degrees from the Jadavpur University. He has more than 35 years of experience in different fields of geophysics with the Geological Survey of India (GSI) at various capacities, and he retired from the GSI as the director (geophysics). Prior to his service in the GSI, Dr Naskar also worked as Chemical and Metallurgical Assistant (CMA) in the Ministry of Railway, government of India, teacher of physics, and an electrical assistant in the government of West Bengal, India. His experience in geophysical mapping employing gravity, magnetics, deep resistivity, GPS, and magnetotelluric (MT) and seismic techniques at different regions of the Geological Survey of India (GSI) is well recognised. He has published more than 100 research publications in recognized national and international journals, as well as book chapters, seminars, workshops, and 31 GSI unpublished reports of mineral exploration, structural and tectonic studies, groundwater investigations, environmental studies, and geophysical mapping in different parts of the country;

currently, he is serving in the National Centre for Seismology (NCS), Ministry of Earth Sciences (MoES), New Delhi. He successfully supervised several field parties of the Geological Survey of India, worked in detailed mapping and a national geophysical mapping programme (in regional and semi-regional scale). Dr. Naskar has enough exposure to interact with geophysicists and scientists of abroad during his services in India; he is a life member of several professional bodies, including Indian geophysical union (IGU), and association of exploration geophysics (AEG).

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1 Geophysical Methods

Fundamentals, Applications, and Case Studies

1.1 INTRODUCTION

The keyword ‘geophysics’ implies the study of the entire Earth system, consisting of the geosphere, the hydrosphere, the cryosphere, and the atmosphere, which allows study in detail by applying the principles of physical sciences based on an assessment of physical property contrasts between the host medium that contains the causative or anomalous bodies and the ambient or surrounding media at depths.

Subsurface geology is explored by geophysical surveys in the field. The geophysical problems are non-linear in nature, and it requires inverse modelling for assimilating a model from the observational data acquired from different sites or observational stations (that means from **data** to **model** [**DM**]), whilst the resolution of the assimilated inversion-based geophysical models is testable by applying the concept of forward modelling, which involves the derivation of data from the assimilated inverted model (that means from **model** to **data** [**MD**]). DM can be best exemplified by citing the exercise of conducting geophysical surveys to collect data from the field to assimilate the model as closer as possible with reference to the true target body or the earth model. In contrary to DM, MD can be exemplified by conducting digitization of the already assimilated existing map to generate data at every grid with systematic distribution, which is regarded as one of the processes of the forward modelling. It is important to mention that the discipline of geology and geophysics are closely related to each other, in the sense that deeper geological structures require geophysical tools for their detection and delineation at different sub-surface depths while solid grasp of geology is necessary for plausible interpretation of geophysical signals that translates into geological details of the surveyed areas. Thus, geophysics is an important tool for geoscientists as well as the entire Earth system scientists to study various aspects of the Earth system. Geophysical experimental studies of subsurface rock samples also used to be carried out in well-equipped geophysical laboratories. The applications of geophysical methods, data generation, processing, and interpretation provide a deep insight into the Earth’s structure at various scales of geophysical exploration, from local to global perspectives, established through rigorous but integrated research on Earth’s structure, its composition, shape, and size.

In this book, we emphasize a variety of geophysical data acquisition field survey techniques to show how different geophysical methods consisting of surface, airborne, shipborne or marine, borehole, or well logging geophysical surveys are being deployed in the field to resolve the issues of metal-bearing minerals, coal, water, oil and gas explorations, and engineering geophysical problems.

The advent of sophisticated geophysical instrumentations and high-speed computational technology in recent years has made sufficient breakthroughs in geophysical exploration survey methodologies to acquire high-quality geophysical data, fast data processing, and robust interpretation of geophysical signals into a reliable and plausible geology of the area under study, which, in turn, has opened ample avenues for generating high-quality huge dataset and their processing in a minimum span of time to meet the requirements of different stakeholders. Nonetheless, the endeavour of geophysical exploration surveys adds significantly to the prosperity of a nation by not only deciphering the hidden reserves of minerals and metal-bearing zones but also resolving intricate issues of infra-structural developments in the areas that are inaccessible, geologically complex, and topographically hostile, through a set of integrated geophysical survey-based solutions.

TABLE 1.1**Methods of Geophysical Exploration Surveys and Applications**

Surface Geophysical Methods	Borehole Geophysical Methods	Airborne Geophysical Methods	Marine Geophysical Methods
• Potential field method (gravity and magnetic) • Electrical and electromagnetic methods (VLF, resistivity, GPR, IP, SP) • Magnetotelluric (MT) • Seismic and Earth imaging methods	• Self-potential (SP) • Resistivity • Nuclear logging • Density logging • Sonic logging • Acoustic logging	• Airborne gravity methods • Airborne magnetic methods • Airborne electromagnetic (EM) method • Airborne spectral method • Remote sensing method	• Seaborne geophysical survey • Seaborne magnetic methods • Seaborne bathymetric survey
Applications	Applications	Applications	Applications
• Resource exploration • Engineering problems and underground structures • Earth physics, structural process	• Well log analysis • Lithostratigraphy and geological formations • Resource quantification • Geotechnical engineering parametric determination • Resource zone identification	• Reconnaissance survey • Detection and delineation of trend of resource deposits • Survey of processes and hazardous zones • Environmental protection measures and resource estimation	• Bathymetric surveys • Deep-sea mineral reserves • Polymetallic nodules • Blue economy

Depending on the applicability of the methodology and the nature of the problem, certain types of suitable geophysical methods are being chosen for their deployment in the field so that geophysical surveys can be carried out purposefully. In view of this, the following flow diagram (Table 1.1) shows how there exist different geophysical methods that are in use in the field, with their varying applications to address specific problems.

It is evident that a single geophysical method is not likely to provide an authentic solution of the problem, but applications of integrated geophysical surveys provide a deep insight to the geological problems to seek geophysical solutions with the economic viability of the geophysical surveys. The results of some of the preliminary geophysical data of the Indian continent, as well as other case histories of different countries, have been discussed, and it has been found that more extensive surveys are an absolute necessity. The geophysical data have contributed to improving the knowledge of the geological structure, which could result in a better understanding of the regional geology and aptness to its mineral potential. Geophysical techniques measure physical phenomena to provide information about the properties of the Earth. Geophysical parameters estimated from the field observation data are generally referred to as *geophysical anomalies*, with reference to an ambience and background. In a lucid language, we can refer anomalies as the difference between the abnormally observed and normally expected. This may be negative or positive, depending upon the values of normal and abnormal with respect to observed and expected, which can be exemplified if one observes a total of 13 fingers in both hands of a person instead of a total of ten fingers, then anomaly is asessed by a difference of +3, and reverse is alos true if observation of total fingers becomes seven instead of ten then anomaly is assessed as −3. Quite often, we use a combination of methods to target the different physical properties of the subsurface and increase the chance of success. Gravity, magnetic, electrical, and magnetotelluric (MT) surveys can be used to locate the rocks and structures which host mineralization.

This chapter gives the readers a brief introduction to the various methods of geophysical exploration in pursuit of exploring *Mother Earth* and its environment. The application of knowledge and

techniques of physics, mathematics, and chemistry to understand the structure and dynamic behaviour of the Earth and its environment has been discussed crisply.

1.2 BROAD DESCRIPTION OF EARTH EXPLORATION

By the beginning of the 20th century, most of the Earth's surface had been explored, at least superficially, using controlled-source geophysical methods, and deeper subsurface had been explored by applying high-resolution geotomography and uncontrollable magnetotelluric methods (Singh et al., 2012; Mishra, 2014), except for the Arctic and Antarctic regions (Apollo 17 mission, 1972). However, polar seismology has proved to be a potential tool for exploring polar regions of the world, including the Third Pole, the Himalayas (Mishra, 2014; Khare et al., 2018). The advent of space science has revolutionized the applicability of geophysical surveys in assimilating quality outputs, and space technology added an advantage to modern geophysical exploration endeavours by positioning itself as the third eye of corroborating surface geophysical results with airborne geophysical data. There is a set of evidence in support of geophysical mapping of each and every inch of land areas that resulted in good maps with the help of radar technology and photographic mapping from aircraft and satellites. It has been reported that the last area to be mapped was the Darien peninsula between the Panama Canal and Colombia (Richard and Harold, 1973). At the same time, geophysical surveys have several constraints that hinder the smooth conduction of geophysical investigations, which include heavy clouds, steady rain, and dense jungle vegetation, which used to make geophysical exploration on the surface very difficult, but airborne radar is able to penetrate cloud covers to produce reliable, detailed maps of the area. Recently, several significant discoveries (e.g. the hidden or abandoned drainage system of the Sahara) have been made on the observation data based on Earth satellites, suggesting the fact related to the relics of a period when the Sahara was not arid. The unravelling of huge mineral reserves in Rajasthan, India, through 'Operation Hardrock' using airborne geophysical surveys (Tajale, 2009); the exploration of *lunar* and *Mars* surfaces using space geophysics; and the discovery of Indian Ocean Geoid Low (IOGL) (Pandey et al., 2022; Negi et al., 2022; Kumar et al., 2022) and hydrothermal fluid zones using shipborne geophysical surveys in the Indian Ocean (NCPOR-MoES, 2018) are some of the significant achievements of recent years that provide impressive examples on the efficacies of geophysical surveys that have explored the impossible targets, which are beyond human reach.

Since World War II, there has been a recorded revolution in the field of geophysical exploration, reflecting the significance of applied geophysics, which unravelled not only the exploration of near-surface but also the hidden mystery of the subsurface Earth to peep inside up to its *core*. It has become an important frontier of studying the composition and thermal regime of the deeper Earth by setting up a diverse set of geophysical advanced observation systems, because solid Earth geophysics has deep probing potential. In recent decades, the exploration of space and deeper ocean has become easier due to the advancement in geophysical sensors, survey methodology, and big data analytics. *Deep Ocean Mission* has become a cutting-edge exploration exercise for any country, including India, to harness natural resources in promoting the *blue* economy of the country. Geophysical exploration allows suitable sites for conducting deep drilling of the upper crust, as witnessed by the deepest boreholes to a depth of 10 km, and India successfully drilled up to a 3 km depth in the Koyna–Warna region of Maharashtra, India, by the Borehole Geophysical Research Laboratory (BGRL) of the National Centre for Seismology, Ministry of Earth Sciences, Government of India (HK Gupta).

1.3 PRIMARY OBJECTIVES AND ACCOMPLISHMENTS

It is a great challenge, in a sense, how the subsurface exploration of hidden resources (e.g. water, fuel, mineral reserves) is possible using geophysical exploration techniques that can contribute to the economic growth and standard of living of the region, which in turn ensures economic profit

and economic incentives to the people. Profit-motivated exploration using advanced geophysical techniques is the need of the hour.

Geophysical investigations of the subsurface are directed to locate the following:

- Oil, natural gas, and coal
- Significant mineral reserves of ores of iron, copper, and uranium
- Reserves of building materials, such as sand, gravel
- Recoverable groundwater
- Geophysical inputs, engineering design and structures
- Geothermal reserves for heating and electricity
- Archaeological features

Geophysics has an immense role in de-risking the solid Earth hazards (earthquakes, landslides, volcanism, tsunamis, avalanches) by providing seismological and geophysical inputs for the development of risk-resilient structures. Sites for dams, power plants, nuclear reactors, factories, tunnels, roads, hazardous waste depositories, and so forth need to be stable and provide assurance that underlying formations will not shift or slide from the weight of the construction, move along a fault during an earthquake, or permit the seepage of water or wastes. Several countries like the USA, Japan, China, and Russia, including India (Bharat), are actively indulged in the precursory study of earthquakes and volcanism because *geophysical* methodology is more reliable in furnishing comprehensive information, since information derived from drilled boreholes cannot provide a piece of complete information alone, rather than information obtained from the boreholes helping verify the geophysical inferences based on subsurface geophysical mapping and imaging at shallow depths.

1.4 METHODOLOGY AND INSTRUMENTATION

Geophysical techniques involve measuring gravity, magnetism, acoustic waves, electricity, electromagnetism, radioactivity, reflectivity, and heat flow. Geophysical equipment under various methods is deployed on the land surface or subsurface under the *surface* and *borehole* geophysical methods; similarly, equipment can be deployed for shipborne in the sea for marine geophysical surveys, as well as mounted to aircraft or satellites for *airborne* or *spaceborne* geophysical surveys. Geophysical methods take care of different physical property contrasts between the causative bodies and the ambience or surroundings beneath the surface. In a real sense, we detect and delineate the plausible geological structures associated with the target reserves of mineral and non-mineral deposits that correspond to the differences in the estimated values of physical property contrasts measured by different geophysical methods. As examples, oil or gas resources associated with sedimentary traps and groundwater associated with sedimentary aquifers, in addition to the presence of dykes, sills, horst, graben, and other host rocks, are to be identified through geophysical characterizations, where minerals or natural resources are found to be concentrated in the subsurface. Different methods depend on different physical properties. The deployment of specific geophysical methods depends on the objective of the survey planned to investigate the types of resources. However, for a greater certainty in deciphering the target bodies beneath the subsurface, required is the application of integrated geophysical methods rather than a single geophysical method, which allows detailed interpretation of different geophysical signals in geology and geological structures.

1.4.1 GRAVITY METHOD

Earth's gravity field in first-order approximation manifests an oblate spheroid. The gravity method is based on the estimation of density contrasts as the gravity anomalies within various strata of the Earth. Usually, surface-based gravimeters are used to measure relative variations in the gravity field at different points. Interestingly, after applying several corrections to the instrumentally

recorded gravity readings/data, gravity anomalies are computed by subtracting a regional field from the measured field to generate residual, representing gravitational anomalies that corroborate with density variations of causative source bodies with respect to the surroundings. Positive gravity anomalies are associated with shallow high-density bodies or subsurface materials of massive crystalline/stronger bodies, whereas gravity lows are associated with shallow low-density bodies or subsurface mass-deficient, weaker/less competent bodies. It is observed that the gravity method is very sensitive to the variation of the density of the mineral bearing rocks, which can be exemplified when the gravity anomalies, if recorded over mapped subsurface structures, are very high, which correspond to the denser and heavy minerals, namely, chromite-, hematite-, and barite-bearing deposits, whilst the low-gravity anomalous signatures over the structures suggest the presence of low-density-bearing structures, such as halite and weathered kimberlite. It is also reported that diatomaceous Earth exhibits gravity lows. Based on the delineation of 'gravity anomalous' zones, a prediction of the total anomalous mass (ore tonnage) can also be made using gravity surveys of the area under investigation. It is important to note that the potential geophysical methods consist of the gravity and magnetic measurements to detect variability in the lateral contrasts in density and magnetic susceptibility, respectively, whilst electrical and seismic methods can detect respective contrasts of resistivity and velocity both laterally and vertically. A very large-scale variation (~1,000 km) can be related to the lateral heterogeneity of the Earth's crust and mantle, and to the thermal and compositional variations. Interestingly, the variation in the gravity field at a very small wavelength, say, lesser than 1 km, may be associated with very shallow anomalies for some specific structures, like a sulphide-bearing formation or a subsurface cavity/void, besides the general topography of the surface.

The gravity method finds diverse applications not only in identifying various mineral deposits but also in finding very useful tools for the identification of lithologies, different geological structures associated with ore bodies themselves (Wright, 1981). Recently, it has been observed that underground coal workings associated with small gravity anomalies can be detectable by combined surface microgravity (10^{-6} Gal) with gravity gradiometric or gravity-gradient metric or tower gravimetric surveys unless they are at shallow depth (Seguin et al., 1994).

Recent studies highlighted the fact that the gravity method opened up widening areas of applications because of unprecedented technological improvements in the design for improving the sensitivity of gravity-measuring devices. High-resolution temporal and spatial variations in the subsurface gravitational field using the gravity method provided a new methodology of gravitation mapping called *4-D gravity survey*. It is also a new dimensional development in gravimetry, where very sensitive satellite-based gravity surveys are being made for monitoring the global field to assess the mass movements associated with hydrology and large-scale glaciers and ice sheets, which may contribute significantly towards the assimilation of a holistic model for the global climate change having local and regional impacts.

1.4.2 MAGNETIC METHOD

The magnetic method exploits small variations in magnetic mineralogy (magnetic iron and iron-titanium oxide minerals, including magnetite, titanomagnetite, and titanohematite, and some iron sulphide minerals, including pyrrhotite and greigite) among rocks. Magnetic surveys deploy different types of measuring equipment, such as fluxgate, proton precession, Overhauser, and optical absorption magnetometers. Generally, we measure the total magnetic field; however, vector measurements are made in some instances. The cause of the changes in the Earth's primary magnetic field in the rocks are associated with various combinations of induced and remnant magnetizations extensively studied by Reynolds et al. (1990). The origin of magnetic anomalies is associated with both induced and remnant magnetization, which are dictated by the quantity, composition, and size of magnetic mineral grains. Primary igneous and sedimentary processes are responsible for magnetic anomalies because these processes involve the genesis of magnetic minerals, a magnetic

mineral alteration that either induces magnetic minerals to the ore body or removes magnetic minerals through several geochemical processes, such as hydrothermal (Hanna, 1969; Criss and Champion, 1984). It is reported that fossil hydrothermal activity can also be detected and delineated by the magnetic surveys. Because rock alteration can affect a change in bulk density as well as magnetization, magnetic anomalies, when corrected for magnetization direction, sometimes coincide with gravity anomalies.

Detection of some iron ore deposits (magnetite or banded iron formation) can be directly made using magnetic exploration. Additionally, magnetic methods are also useful for deducing subsurface lithology and structure that may indirectly corroborate with the location of mineralized rock, patterns of effluent flow, and extent of permissive terranes and/or favourable tracts for deposits beneath the surficial cover. The identification of magnetic minerals associated with ore or waste rock from which hazardous materials can be deciphered by geo-environmental applications of the magnetic methods in turn provides indirect identification of hazardous materials associated with many nickel-copper or serpentine-hosted asbestos deposits, since all these rocks lose their magnetic properties beyond a certain temperature, called the Curie temperature (500–600°C for peridotite).

1.4.3 SEISMIC METHOD

In order to investigate subterranean structures related to the exploration for petroleum, natural gas, and mineral deposits and the shallow engineering problems, seismic controlled-source method is found as the most reliable and very apt method of exploration, not only for resource identification, but also for delineation of causative factors where seismogenic strengths of the subsurface rock materials get changed. The methodology is based on the assessment of time elapsed in travelling of the seismic wave between the selected shot point and the arrival of reflected or refracted impulses at one or more seismic detectors. Several sources, especially seismic airguns, are commonly used to initiate the seismic waves. The common example for seismic source is to generate sound wave through hammering the iron or steel plate kept on the surface. Several sufficient blows on the surface either by thumper or vibrosis or through blows on iron or steel plate by the sledge hammer of the specific weight generate sound wave that impinges into the ground and travels in the subsurface layers. Interestingly, the speed of sound wave impinged into the ground becomes the speed of elastic wave as the propagation of the wave gets dictated by the elastic moduli of the sub-surface rock materials. This technique has now been replaced by the source generated by exploding dynamite underground. Presently, vibroseis and electric vibrators or falling weights (thumpers) may also be employed at sites to generate seismic energy for conducting seismic exploration surveys. It is a great caution that an underground explosion might cause damage once the surface is associated with cracked rocks or once the locations where blasting is to be made get associated with caverns. The seismic surveys generate seismic traces or seismograms as the record of the arrival times of the seismic wave by the detectors with their varying amplitudes, representing the record of the ground vibrations. Two processes are involved in seismic exploration that constitute two methodologies in seismic exploration; one is the *reflection* seismic method, which provides fine structural detail, and the *refraction* method provides precise estimates of depth to lithologies of differing acoustic impedance. Mineral investigations to map low-velocity alluvial deposits such as those that may contain gold, tin, or sand and gravel are made by deploying seismic refraction surveys.

1.4.4 ELECTRICAL METHOD

Based on the deployment of different instruments and procedures, the *electrical* methods comprise a multiplicity of separate techniques which have variable exploration depth and lateral resolution. Several acronyms describe techniques and their variants. Electrical methods can be described in five classes: direct current resistivity, electromagnetic, mise-a-la-masse, induced polarization, and self-potential. It is observed that despite several variants, the fundamental measurements of the

Earth's field electrical impedance relate to changes in impedance, and also to measure changes in electrical potential with respect to the given flow of current in the media. As mentioned earlier, electrical methods also find a wide-scale applicability for mineral and geo-environmental problems. The electrical methods are used to identify sulphide-bearing minerals. This methodology can also be directly applicable to groundwater exploration and also be used to identify subsurface structures and lithologies.

1.4.4.1 Direct Current Resistivity Method

It is evident from the law of resistivity that conductivity is the inverse of the resistivity of the media wherein the direct current resistivity methods can be deployed to measure Earth's resistivity using a direct or low-frequency alternating current source. It is a well-documented fact that rocks are electrically conductive, and due to the ionic mobility that takes place in pore spaces associated with water, besides electronic conduction through metallic lustrous minerals, such phenomenon is not common. It is so because of the metallic lustrous minerals devoid of providing long continuous circuit paths that require for conduction in the host rock. Bulk rock resistivity plays an important role since it is associated with water content and dissolved ionic materials of the media in which the current flows. As per the structural and morphological settings of the rock, the high porosity contributes towards the low resistivity in water-saturated rocks. Several case studies show direct current techniques applied to a variety of mineral exploration and geo-environmental considerations with reference to various ore deposits. It is observed that the low resistivity enables the detection and delineation of massive sulphide deposits. Another important evidence of low-resistivity target is the clay alteration assemblage in and around the hydrothermal systems. Identification of lithologies and structures that control mineralization is made by observing the wide range of Earth material resistivities using electrical resistivity. Additionally, the electrical resistivity methodology is applicable to deciphering acid mine waste because the high mobility of the hydrogen ion makes the target conductive in comparison to the solutions that contain equivalent concentrations of neutral salts in the subsurface layers.

1.4.4.2 Electromagnetic Method

It is a proven fact that the electromagnetic geophysical method is based on the use of alternating magnetic fields to induce a measurable current in the Earth. Mineral exploration for detecting and delineating low-resistive or high-conductive massive sulphide deposits has been carried out by deploying electromagnetic (EM) methods. EM-based airborne methods are generally used for scanning larger surface areas for investigations that provide a multitude of targets for ground surveys. The methodology includes an airborne electromagnetic survey that found a wide-scale application in mapping lithologic and structural features (Hoover et al., 1991) that may help for various mineral exploration and in addressing geo-environmental problems.

1.4.4.3 Mise-a-la-Masse Method

A state-of-the-art geophysical method applied to conductive masses having larger resistivity contrasts with their surrounding host rock is referred to as the *mise-a-la-masse* method. The *mise-a-la-masse* method has found its recognition in mineral exploration in mapping massive sulphide deposits. There are several successful case studies for geo-environmental investigations of highly conductive targets for which contaminant plumes originating from abandoned mines are identified (Osiensky and Donaldson, 1994). Mineral belts of Agucha and Dariba of Rajasthan, India, were investigated using the geophysical methods by the Geological Survey of India (Pant and Gupta, 1998; Mathur et al., 2003; Gupta, 2007; Gopal et al., 2008). The *mise-a-la-masse* method provides a very reliable inference on sulphide deposits beneath the Agucha and Dariba mineral belts of Rajasthan, Bharat (India), which is found well corroborated with other surface geophysical and airborne geophysical anomalies estimated independently (Gupta, 2007).

1.4.4.4 Self-Potential Method

Natural sources are akin to generate appreciable direct current or quasi-direct current associated with natural electrical fields of self-potentials. Generally, sulphide mineral deposits exhibit self-potential anomaly through the processes of ongoing oxidation (Mathur et al., 2003; Gopal et al., 2008), and when metals are mobilized, alternatively, self-potential anomalies also get originated due to fluxes of water or heat through the Earth (Corwin, 1990). The self-potential method gets applications addressing issues related to geo-environmental ones in searching of zones of oxidation and passage of groundwater in rock matrices. It also has applicability in deciphering mine fire subsurface zones in the Raniganj Coalfield area of West Bengal, India, which were investigated by the Geological Survey of India (Biswas et al., 1999; Ghosh and Mishra, 2000).

1.4.4.5 Induced Polarization Method

Polarizable minerals, such as metallic-lustre sulphide minerals, clays, and zeolites within water-bearing pore spaces of rocks, are detected and delineated by deploying the induced polarization method, which gives responses to active surface areas within rocks. It is observed that disseminated sulphide minerals are well detectable in comparison to that of massive sulphide deposits. The principal cause of generating measurable induced polarization in massive sulphide deposits is the presence of significant gangue in the deposited sulphide bodies. Induced polarization has found its greatest application in the exploration for disseminated sulphide ore, where it may detect as little as 0.5% (v/v) total metallic lustre sulphide minerals (Sumner, 1976). In geo-environmental studies, induced polarization surveys are principally used to identify sulphide minerals, but it may have other applications, such as outlining clay aquitards that can control mine effluent flow.

1.4.5 RADIOACTIVE METHOD

When rocks or ore bodies are associated with radioactive minerals beneath the surface, that warrants the deployment of the radioactive method for their detection and delineation. Natural radioactive materials exist in the upper part of the crust and are associated with uranium, thorium, and a radioisotope of potassium (potassium-40), as well as from radon gas. Spontaneous emissions of alpha and beta particles, along with gamma rays, during the disintegration and decay processes of the radioactive materials provide us with an opportunity to detect radioactive rocks and minerals at the subsurface. Detection is usually of gamma rays, and it is accomplished in most cases with a scintillometer, a photo-conversion device containing a crystal of sodium iodide that emits a packet of electromagnetic radiation, called photon (e.g. packet of energy), that is generated when struck by a gamma ray whose energy is determined by the nature extent of the radioactive disintegration or decaying ability of the radioactive materials that provide a way forward to estimate electron counts using a photocathode device in the measuring equipment.

1.4.6 GAMMA RAY METHOD

In recent decades, several geophysicists have demonstrated that the gamma ray method is a very apt geophysical methodology that uses a scintillator to identify the presence of natural radioelements, such as potassium, uranium, and thorium, along with the measurement of individual radioelement abundances, with the help of multichannel spectrometry (Durrance, 1986; Hoover et al., 1991). This method finds a wide-scale application in uranium exploration because of the direct detection of the radioelements. Thorium with felsic rocks exhibits similar behaviour like that of zirconium because the immobility of the radioelements is easily detected by the gamma ray geophysical method, which also provides geo-environmental information concerning radiation dose and radon potential. Some of the deposits in or adjacent get enriched in uranium and/or potassium, which helps indirectly to assess the potential for the release of hazardous materials from ore or waste piles. Additionally, the

oxidation of sulphide minerals in the radiogenic rock matrix accelerates uranium mobilization for easy detection using the gamma ray method.

1.4.7 REMOTE SENSING METHOD

The remote sensing method is based on measurements of electromagnetic radiation from the ground of reflected energy falling in various spectral ranges by sensors mounted to aircraft and satellite. It is a fact that remote sensing encompasses aerial photography and other kinds of measurements that are generally displayed in the form of photograph-like images. This methodology has applications involving a broad range of studies when the area of investigation is inaccessible and located in a hostile, undulatory terrain, including cartographic, botanical, geological, and military investigations.

1.4.8 GEOTHERMAL METHOD

Heat flow anomalies are detected indirectly by the temperature gradient measurements to ascertain geothermal resources, such as sulphurated water and steam. Deployment of resistivity, seismic, and magnetotelluric geophysical exploration techniques helps in mapping the geothermal reservoirs located at few hundred meters beneath the surface because of changes in the electrical and elastic properties due to appreciable variation in temperature. Volcanic areas around the world are associated with significant temperature changes in their magma chambers, making them excellent sources of geothermal energy as an alternate resource for energy security due to geothermic processes. In India, two volcanic zones located at Barren and Narcondum in the Andaman-Nicobar region witnessed their sporadic eruptions, indicating the presence of potential resource zones for geothermal energy in India. The hot spring zones indicate the presence of subsurface shear zones and faults associated with various geological structures that control the flow of hot underground water, which are very much detectable using suitable geophysical surveys.

1.4.9 GEOCHEMICAL METHOD

It is a documented fact that a set of chemical elements is transported in varying quantities, from minute to small, by fluids flowing into the Earth, which allow systematic measurement of such trace elements, which in turn helps in locating and detecting their sources. Geochemical methods are very sensitive in the sense of the applications of several chemical analyses that provide reliable detection and estimation of minute amounts of materials in the subsurface rock matrix and in the mineral assemblage. It is reported that trace elements are used for the distinct identification of the specific category of hydrocarbons present beneath the surface. The hydrocarbons may be of different categories, such as petroleum, natural gas, and other fossil fuels. Industrial wastes, along with the identification of soil contamination and sites of pollution discharge into rivers, are being explored through geochemical soil mapping of smaller areas, as well as of the entire country, through suitable sampling of the targeted areas.

1.4.10 EXCAVATION, BORING, AND SAMPLING

To make positive identification of ores, fuels, and other materials, direct sampling by means of boreholes is required, which helps in quantifying the amount of recovery through the applications of geophysical measurements. Boreholes can be drilled in both hard and soft rocks by deploying suitable drilling machines. After a borehole has been drilled, various tools—sondes—are lowered into the hole to measure different physical properties using geophysical well logging techniques. Several boreholes have been drilled and geophysically logged for exploratory purposes in the state of Assam by Oil India Limited (OIL), one of the Maharatna companies of Government of India. The borehole depths are found to vary from 1,600 m to over 5,000 m in Assam region of India in

which seismographic sensors can be installed for imaging the sub-surface areas using 3-D seismic tomography that provides detailed information about the deeper structures.

Recently, borehole geophysics achieved significant results in unravelling both shallow for deciphering mineral reserves (Gupta, 2006) and the deeper structures of the stable intra-plate region to ascertain earthquake nucleation processes with reference to rock physics and its bearing on seismic hazards associated with the reservoir-triggered earthquakes (Gupta et al., 2016) as witnessed by a series of research publications from the Borehole Geophysical Research Laboratory (BGRL). The National Centre for Seismology with its BGRL ascribed to Ministry of Earth Sciences, India and the National Geophysical Research Institute (NGRI), CSIR, India enabled, for the first time, to drill a borehole 3 km deep for conducting pilot studies that would be extended further to a 7 km depth for conducting integrated seismo-geophysical-geotechnical-bio-geophysical-geological studies to address the intricacies of the seismotectonic strata beneath the Koyna–Warna region (Gupta et al., 2016; Roy et al., 2013; Roy, 2017).

1.5 GEOLOGICAL AND GEOPHYSICAL INTEGRATED INTERPRETATION AND MODELLING TECHNIQUES

Integrated and state-of-the-art mineral exploration aims of targeting mineral resources at depth or under cover. This requires moving beyond traditional interpretation of data and its map representation towards interpretation of 3-D models for which assimilated model is used to provide interpretation into an actionable construct to adopt a decisive planning for mineral mining. Integrated interpretation allows the application of multidisciplinary data integration directly in terms of exploration for awareness in the industry (Joly et al., 2015; McGaughey, 2006; Mitchinson et al., 2014; Chalke and McGaughey, 2015). Such approach minimizes ambiguity and provides answers to geoscientific questions that are more reliable. Nonetheless, integrated interpretation is also significant for the development of the geological models by integrating geophysical data with the available geological information for the targeted area (Jessell et al., 2014; Lindsay et al., 2013; Spampinato et al., 2015) that is a part of the multidisciplinary targeting processes, since we are always poised to translate geophysical data/signals into the geology. Generally, 3-D geological models have predominantly been utilized in a mining environment where interpretation is facilitated by large volumes of information obtained through dense drilling and pit/underground mapping. It is a fact that 3-D Earth models have been based primarily on geophysical inversion, which requires several geological information, such as smooth distribution of rock properties, geological setups, and petrophysical parameters (Li and Oldenburg, 1996), so that the model can be interpreted in terms of geological domains and get validated against the geophysical surveys.

Based on the availability of the data and geological objective, a plausible but suitable methodology for integrated interpretation of geological and geophysical data is chosen. It is so because data-rich environments consider the integrated interpretation of geological and geophysical data for constructing a geological model with geological constraints; then, corroboration of the model with petrophysical data is made, and finally, validation and reconciliation of a model are obtained through forward modelling and inversion. This is essentially a model validation but does still reveal new information and advance the understanding of the relationship between geology and rock properties.

In other cases, where subsurface control is limited, a geologically based model can be developed through 3-D investigative modelling of geophysical data and other datasets, since an unconstrained inversion is not much reliable but rather largely conceptual. Interpretation of 3-D geological models derived from integrated geological and geophysical data helps in understanding the relationship between geology, geophysical datasets, and rock properties. There are several key steps to be followed for assimilating an integrated geological and geophysical modelling:

- Interpret the available data. Assess the geophysical signatures and their relationship with geology and rock properties.

- Develop a starting model and establish a geological framework for modelling.
- Adoption of quantitative modelling and inversion with reference to conceptual ideas to incorporate into the model.
- Model validation and update using geophysical inversion techniques in close collaboration with 3-D geological modelling.

This ‘geological framework’ is not always clearly defined at the start of the project. Often, 3-D modelling is hindered by attempting to incorporate too much detail into the geological model that the geophysical survey is simply not responding to the targeted problems. In order to identify the key domains to incorporate into the geological framework, quantitative forward modelling is required, despite the constraints mapping or drilling provides, especially showing how the geophysical responses are responding to known geological domains. In the absence of subsurface constraints, it can be conceptual. Though both forward modelling and inversion are applied comprehensively to geophysical exploration by integrating geological and geophysical data, the ease and speed of completing a forward model of geologically based problems provide a direct means for testing and validating geological hypotheses. Furthermore, inversion options (adjustment of properties and/or geological boundaries) provide feedback mechanisms for updating the geological model (either directly or indirectly). Several algorithms on inversion are available for shallow subsurface layer investigations (e.g. VPmg, VPem1-D, VPem3-D) (Fullagar and Pears, 2007). These work directly on geological models, with the option of adjusting model geometry or model properties for both homogeneous and heterogeneous conditions as a key driver for integrated interpretation (Pears et al., 2001; Fullagar et al., 2008, 2010; Fullagar et al., 2013).

The selection of a geological framework is of paramount importance because it permits for the conduction of several hit and trial tests with adjustment using forward modelling and inversion techniques to achieve a geological model which should be consistent with the measured geophysical responses. Forward modelling and observing a good correlation between the predicted and measured data give insight into the robustness of the geological domains, which gets validated by assigning a homogeneous physical property to each domain; a final stage of inversion is to solve for local property variations within the domains that highlight anomalies, which may be associated with potential targets or areas of geological complexity that require further investigations.

1.5.1 SOME SUCCESSFUL CASE STUDIES OF GEOPHYSICAL INTERPRETATION

There are several successful case studies available for geophysical interpretation to ascertain the bearing of geophysical signals into geology and geotectonics. Case studies use data from the regionally extensive magnetite-bearing sedimentary strata within the area of study, such as Minotaur’s Mutooroo project area in South Australia, with the objective of measuring magnetic susceptibility-based magnetite percentage in the host rock, which has been found difficult to directly equate to magnetite percentage and prompted for a geologically based approach to modelling. The adopted methodology is therefore to (a) define the geometry of the primary magnetic domains and then to (b) invert for local magnetic susceptibility variations within those domains. A total of seven boreholes were drilled for control, and a geologically based 3-D interpretation through modelling of aeromagnetic data was developed (Pears and Chalke, 2019). The aeromagnetic survey was flown EW with 50 m line spacing and a flight height of 25 m. Euler deconvolution depth to source techniques was applied to ascertain the base of weathering. Depth was computed using the borehole data.

Modelling of discrete magnetic domains was based on interpreting the subcropping signatures of these domains from the 1VD RTP (reduction to pole), then assessing forward-modelled responses for 3-D bodies. An optimal dip for the magnetic signatures associated with the various magnetic domains across the area was determined using the concept of the dipole nature of the magnetic responses (Pears and Chalke, 2019). The dip that had a computed response that best matched the measured data was incorporated into the full 3-D model. Using this approach, various portions of

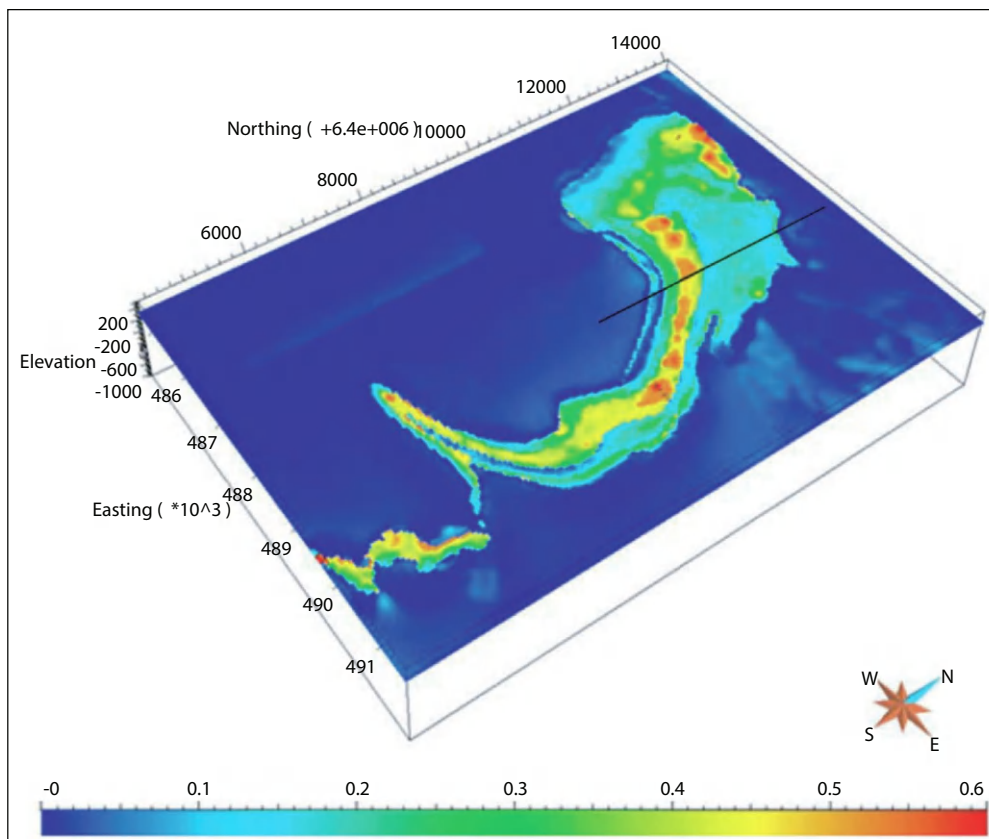


FIGURE 1.1 The final inverted model of the study area in Chile, with 3-D magnetic susceptibility variations within the domains.

the magnetic anomalies across the area were investigated separately, and the geometry from the individual breakout investigations was combined into a final 3-D model. VPMg inversion techniques were used to optimize the homogeneous susceptibility assigned to each domain and locally refine the geometry of modelled domains (in particular, depth of weathering). The final 3-D magnetic domains of the model (beneath cover) are illustrated by Pears and Chalke (2019); each domain is coloured by magnetic susceptibility (Figure 1.1). The computed response of the final 3-D geological model is with homogeneous susceptibility (Pears and Chalke, 2019). In order to solve for local susceptibility variations within the modelled domains, the final stage of inversion is considered to ascertain anomalous zones (Phelps-Barber et al., 2022). High susceptibility indicates high targets (increased magnetite percentage). Figure 1.1 illustrates a horizontal section through the inverted model. The location of the section is illustrated in Figure 1.1.

1.5.2 REGIONAL BASEMENT MODELLING

Regional basement modelling has occupied an important place in the geophysical interpretation that unravels several features of the mineral resources. The modelling procedure may be adopted as follows:

1. Compile all geological and geophysical data into a 3-D modelling software environment, and assess the relationships between drill holes, mapping, and geophysical signatures.

2. Address the prominent regional trend across the gravity data using 3-D modelling techniques.
3. Adopt a simple two-layer starting model.
4. Apply the forward modelling with gravity data to assess the starting model response.
5. Drill hole–constrained gravity geometry inversion to update the basement topography.
6. Validate the updated basement topography: preliminary heterogeneous density inversion in the domains and forward model basement topography against the magnetics data.
7. Apply geophysical inversion of the gravity and magnetic data to assess the basement density and magnetic susceptibility variations with the constraint of the final basement topography.

A very important study on Atacama gravels overlying intermediate intrusive and volcanics in northern Chile provided the best geophysical interpretation to address the geological problems (Pears and Chalke, 2019). The area of interest is largely covered by Atacama gravels overlying intermediate intrusive and volcanics in northern Chile (Pears and Chalke, 2019). Exploration is for Cu-Au porphyry deposits; exploration in similar terranes has been published by Hope and Anderson (2015). In this case study, the aim of the integrated interpretation is, first, to define the depth of the basement, then to utilize the modelled basement contact as a constraint for gravity and magnetic inversion to resolve density and magnetic susceptibility in the basement. It is observed that the VPmg software has very useful applications for the potential field of forward modelling and inversion. The quantitative basement modelling using potential field data follows a similar procedure to that described by Pears et al. (2001). Regional gravity data was collected at 400 m × 400 m station spacing across the project area. The airborne magnetic data was from a survey flown with NNE-orientated flight lines at 100 m spacing. The nominal sensor height for the magnetic survey is 30 m. There were 11 drill holes across the project area, of which only 5 were in the area covered by gravity (Pears and Chalke, 2019).

Based on drilling, expected basement depth is in the order of 100 m to 200 m. Summaries of the density and susceptibility information were provided but were based on limited samples and general knowledge of rock types in the area. A density contrast of 0.6 gm/cc between cover and basement was assumed for modelling basement topography (~2.1 gm/cc overlying 2.7 gm/cc basement). This case study highlighted a basement topography model which is consistent with geology and the potential field data. Basement density and magnetic susceptibility variations are also assessed using the concept of geological. They also assessed inverted basement densities and magnetic susceptibilities and found that there are discrete zones of low density (e.g. felsic intrusive) that were also enveloped by increased magnetic susceptibility (possibly associated with potassic alteration typical of porphyritic-style mineralization). The most spectacular geophysical-derived model unravelled the hidden mystery of the decreased density within the basement and the surrounding zones of increased magnetic susceptibility obtained after querying the targeted zones (Pears and Chalke, 2019).

During the course of the interpretation, potential field modelling revealed that gravity was responding more strongly to the basement topography than do magnetics, and accordingly, the derived basement model was largely dependent on quantitative modelling of the gravity. Subsequent drilling has validated basement depth, strengthening confidence in the modelled basement geometry and heterogeneous properties and drilling results associated with mineralization. Another successful case study of geophysical interpretation is of the Stanthorpe region, where aeromagnetic, gravity, and radiometric geophysical data are used in the genesis of the model (Kumwenda and Lackie, 2019). This study provides an in-depth geophysical interpretation of the geology of the Stanthorpe region (Queensland, Australia), with a particular focus on hydrothermal alteration. The Stanthorpe region is part of the New England Orogen (NEO), a complex tectonic collage of different terranes accreted and interacted with the eastern margin of the Australian continent during the late Palaeozoic to early Mesozoic (Flood and Aitchison, 1993). The study applied from Sydney

to Brisbane and is separated by the Mesozoic cover of the Clarence–Moreton Sedimentary Basin (Mustard et al., 2006).

Airborne geophysical methods include magnetic, radiometric, gravity, and electromagnetic, with the first three techniques mostly applicable in geological mapping (Dentith and Mudge, 2014). These methods, just like other geophysical techniques, utilize certain physical properties, such as magnetization, radioactivity, and density, to map subsurface features. Measurements of these physical parameters are thus used to infer the geology of the survey area (Dentith and Mudge, 2014). Both magnetic and gravity surveys are useful in mapping structural features, such as faults, and delineating geological boundaries. However, gravity surveys have a much greater depth of investigation, and gravity data are often used to map geological features at a deeper level than the aeromagnetic method (Jaques et al., 1997). Airborne radiometric surveys measure variations in three radioelements, namely, potassium, thorium, and uranium, in the soil and exposed rocks in the upper 30 cm of the Earth's surface (Dickson and Scott, 1997). It is worth mentioning that radiometric surveying has a very shallow depth of penetration and investigation in comparison to that of magnetic and gravity geophysical methods deployed for the study, which is the greatest constraint of the radiometric surveying that fails in mapping deep-lying structures. Variations in the concentration of radioelements mainly relate to changes in the mineralogy and geochemistry of rocks and regolith materials (Wilford et al., 2009), because of which lithological changes are easily mapped (Dickson and Scott, 1997).

Therefore, an analysis of the geophysical signatures related to various rock types in the Stanthorpe region can help uncover concealed structures, such as faults, fractures, joints, and dykes; delineate geological boundaries and alteration zones; and more importantly, estimate depths and widths of anomalous areas. In this way, geophysical responses provide a much clearer picture of the geology of the Stanthorpe area. All the geophysical data (magnetic, gravity, and radiometric) used to interpret the geology of Stanthorpe were downloaded using the Geophysical Archive Data Delivery System (GADDS) on the Australian government portal (www.geoscience.gov.au/gadds). Total magnetic intensity (TMI) data have a grid cell spacing of ~80 m, gravity has a grid spacing of ~800 m, and the grid spacing of radiometric data (K, Th, U, and dose) is ~100 m. *Dose channel* is a combination of all three radioelements (K, Th, and U). The interpretation of airborne geophysical data presented by Kumwenda and Lackie (2019) provides an insight into the geology of Stanthorpe (Queensland, Australia), particularly areas affected by hydrothermal activities.

It is observed that the study areas associated with high concentration of faults and dykes exhibit low density and low magnetic susceptibility, which furnishes strong indication for the existence of hydrothermally altered rocks in the areas that found corroborated with radiometric signatures with high potassium (K), low thorium (Th), and low thorium–potassium ratio (Th/K). Modelling of magnetic and gravity data reveals a zone of low magnetic susceptibility that coincides with a section of the gravity profiles having the lowest density values. This study also highlighted that the non-magnetic zone associated with low-density rocks showed a different pattern of anomalies distributed between the north and the south. It is worth bringing the fact to the attention of the readers that the geological map of Stanthorpe has not outlined the altered sections; however, borehole data provided evidence that the Stanthorpe batholith is underlain by substantial volumes of strongly quartz-sericite-chlorite-muscovite-altered granitic rocks. These findings, coupled with highlighted geophysical evidence, suggest the existence of altered rocks at depth and help outline hydrothermal alteration zones despite being not well exposed to the surface. Such inferences invite for further geological research for areas of potential hydrothermal alteration to ascertain the extent of the alteration zones in the Stanthorpe region, and also to know the cause of the alteration thought to be the circulation of hydrothermal fluids in faults, fractures, joints, and veins generated by the emplacement of the Stanthorpe granitoids. Geophysical interpretation of low magnetic response, moderate gravity response, and high Th with low K and low uranium (U) is shown in **Figure 1.2b** and **Figure 1.2c** (Kumwenda and Lackie, 2019). Interestingly, very low density and high K are interpreted as blind plutons. These plutons, however, did not appear on the geology maps of Stanthorpe, and it proves the efficacy of geophysical methods to provide plausible interpretation of the deep geology.

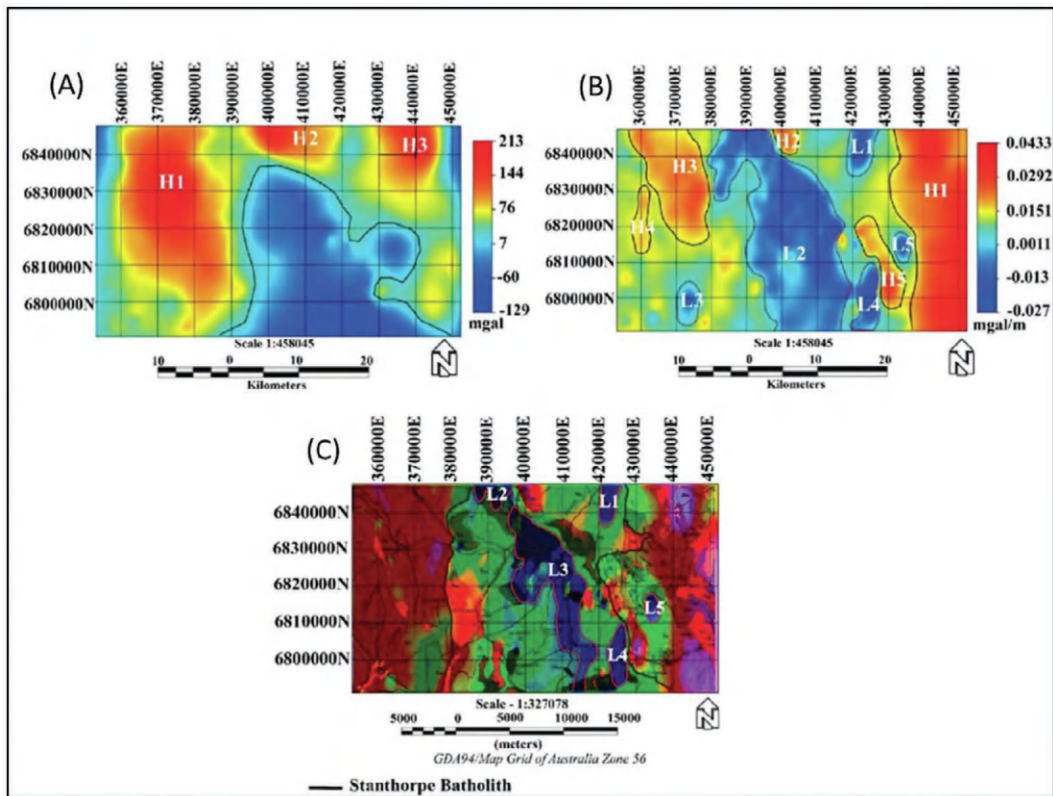


FIGURE 1.2 (a) Residual anomaly map of the Stanthorpe region. (b) IVD anomaly map. (c) An overlay of the geology map of Stanthorpe on the IVD image. Black lines represent the extent of the Stanthorpe batholith extracted from a geology map, and $L1$ to $L5$ are areas with distinct negative values.

Geophysical exploration for gold and associated minerals is well documented for several outstanding geophysical studies carried out in different parts of the world, such as in India, China, Australia, Russia, USA, Canada, Ghana, Indonesia, Peru, Kazakhstan, Mexico, Uzbekistan, South Africa, Sudan, Brazil, Papua New Guinea, Mali, Argentina, Egypt, Meyduk (Latala) in Kerman Province, and several others.

The Wadi El Beida area, South-Eastern Desert, Egypt, provides a very successful case study for geophysical exploration for gold and associated minerals using different techniques, such as magnetic, self-potential (SP), induced polarization (IP), and resistivity, which are adjudged as important geophysical tools for mineral exploration for ores located in basement rocks (Sultan et al., 2009). As described earlier, integrated geophysical methods for mineral exploration are a very reliable approach for geoscientists (Macnae, 1979; Smith, 2002). Several researchers used the IP technique for mineral exploration as well as for hydrological and environmental geology studies (Kiberu, 2002; Vacquier et al., 1957; Marshall and Madden, 1959; Sternberg and Oehler, 1990; Towel et al., 1985; Sumner, 1976; Klein and Sill, 1982). The self-potential (SP) method has a wide range of applications in engineering and geotechnical investigations (Corwin, 1984; Markiewicz et al., 1984), in geothermal exploration (Corwin and Hoover, 1979; Anderson, 1984), and in the exploration for minerals, particularly metallic sulphides (Corry, 1981; Yungul, 1950). The magnetic technique was used for identifying massive sulphide zones (Ramadan and Sultan, 2004; Gupta, 2007). A classical study by Sultan et al. (2009) based on integrated interpretation to estimate the ore bodies that include gold and associated minerals demonstrated that geological,

magnetic, self-potential (SP), gradient resistivity, induced polarization (IP), and dipole–dipole techniques help outline the areas of alteration associated with the gold, and those with mineral deposits, in the Wadi El Beida area, South-Eastern Desert, Egypt (Sultan et al., 2009), since pyritic alteration produces anomalous geophysical signatures as indicative of the gold that cannot be detected directly.

1.5.2.1 Magnetic Data

The total intensity magnetic map (Sultan et al., 2009) reveals that the magnetic anomalies can be grouped into different types, locations, and trends. The north-western part of the area is characterized by very low magnetic anomalies (37,724 nT) of the NW–SE trend corresponding to meta-andesite rocks. The north-eastern part of the area is occupied by high magnetic anomalies (41,041 nT) related to ore bodies; the eastern and southern parts of the area are characterized by moderate magnetic anomalies. The first vertical derivative of the total intensity data suggested the existence of a well-oriented trend of the structural anomalies. Regional magnetic data are being used to find out the basement structures and mineralization (Bangarubabu et al., 2022).

Most importantly, 3-D magnetic inversion using Euler deconvolution and the magnetic models using the MAGMOD program are used for estimating the magnetic depth. A quantitative interpretation of the total intensity magnetic map is carried out by using the GRIDEPTH program, which is part of the Geosoft Program Package (1998), which is used by most geophysicists in the world, including several organizations of India.

1.5.2.2 Self Potential (SP)

Recent advancement in geophysical computational domain improved the quantitative interpretation of self-potential anomalies based on two techniques: the first is a new algorithm constructed by Monteiro Santos (2009) using particle swarm optimization (PSO), and the second is the code constructed by Caglar (2000). Since the SP anomaly produced by the polarized structures along a principal profile over the body (Yungul, 1950) for different shape factors of 1.5, 1.0, and 0.5 that correspond to sphere (3-D), a horizontal cylinder (2-D), and a semi-infinite vertical cylinder (3-D), respectively, a sheet-like body is better diagnosed by the SP anomalies (Murthy and Haricharan, 1984; Atchuta Rao and Ram Babu, 1983). Interestingly, the sheet-like ore body deciphered by geologic observations and magnetic interpretation requires the PSO algorithm to invert the SP data to corroborate and confirm the fact that the detected ore body is represented by a sheet-like body, and its depth can also be well estimated. The SP anomaly is also an indicator of the existence of several geological structures. It is observed that weak SP negative anomalies suggest the presence of high shearing zone in the study area (Sultan et al., 2009). A classical study by Gopal et al. (2008) using SP with the the drill-hole induced polarization and magnetic surveys identified the potential mineralized belts in Rajasthan province of India.

1.5.2.3 Gradient Resistivity Map

The gradient resistivity map, as described by Sultan et al. (2009), reveals a low-resistivity zone related to ore body ranging from 277 to 366 Ωm with its specific trend. The quantitative interpretation of gradient resistivity can be carried out along different profiles, and the resistivity gradient map can be constructed using the Schulz (1985) method. A set of parameters is generally estimated, such as depth to the centre of the body, radius (R), depth to the top, and maximum width of the causative body.

1.5.2.4 Gradient IP Map

A very successful case study conducted for Egypt by Sultan et al. (2009) showed a good correlation with high-chargeability (6–13 mV/V) zone uses ranging from and related to the ore body which is located at the same site of the gradient resistivity map (low-resistivity zone).

1.5.2.5 Dipole–Dipole Sections

Four dipole–dipole profiles were carried out at the northern, central, and eastern parts of the study area (Sultan et al., 2009). The inverse model obtained from the profile suggested two horizontally disposed heterogeneities associated with low resistivity and relatively high chargeability. The first extends from the beginning of the profile up to a distance of about 75 m within a depth range of 15–20 m, while the second ore body is located at a distance of about 105 m in a depth ranging from 15 to 35 m. The inverse model obtained from other profiles at the central part of the area reveals three ore bodies. One of them is relatively deep-seated, at a depth ranging from 25 to 65 m, and extending from a distance of 115 to 160 m. The dipole–dipole section along another profile located at the central part of the study area explained an ore body from the central part extending to a depth ranging from 15 to 40 m. High chargeability and low resistivity at distances from 70 to 120 m and depths of 40–65 m give good correlation.

The gradient resistivity map and IP chargeability map show the same trend of low resistivity and high chargeability, from which the depth of the ore body was estimated at a range of 21.2–62.3 m and a maximum width ranging from 52.6 to 165.2 m (Sultan et al., 2009). Thus, we can infer that the low resistivity and the high chargeability of the dipole–dipole section are well suited with the gradient resistivity and IP maps. The results of mineralogical analyses from two field samples have indicated that the ore bodies, which have been detected from geophysical interpretation, are rich with a gold content of 5–502.5 gm/ton.

From the interpretation of the geological, geophysical, and geochemistry data, it is concluded that gold and disseminated sulphides are located on an alteration shear zone which is composed of quartz-feldspathic. Sultan et al. (2009) showed that the area of study has high gold grades associated with shallow linear features, which are conspicuously identifiable by a number of geophysical techniques (Figure 1.3). Further, they suggested that the plausible dimension/shape/form of any

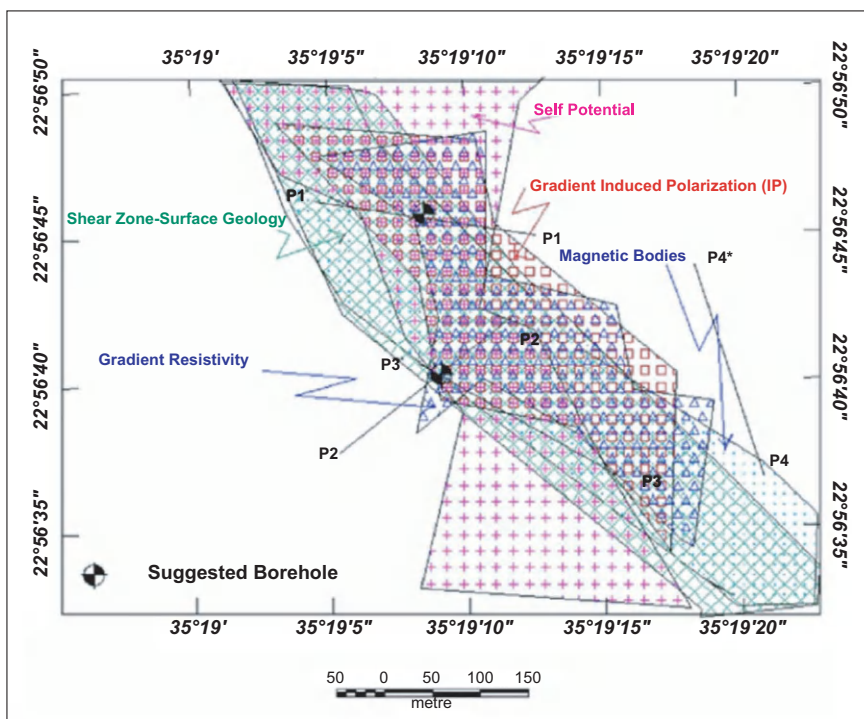


FIGURE 1.3 Integrated interpretation map for the geological and geophysical results of the study area in Egypt.

potential ore body may be confirmed by conducting exploratory borehole drilling at the demarcated and delineated sites identified by geophysical investigations.

1.6 GEOTECHNICAL INTERPRETATION OF GEOPHYSICAL DATA

Geophysical exploration has a proven track record of interpreting geophysical data to address geotechnical and geo-engineering problems and disaster risk resilient issues, including isolated geologic structures (Mishra, 2020; Mishra et al., 2020, 2022). Geotechnical investigations are very much closely related to geophysical surveys and are observed as cost-effective in deciphering shallow information under near-surface geophysics (Sirles, 2006). In order to get reliable information about the nature and variability of the subsurface heterogeneities between the existing boreholes, conducting either a cross-hole or a surface-to-borehole, or both, geophysical survey is deployed, which is referred to as geotomography, which measures the different geophysical parameters of the area under study (Seguin et al., 1994). An isolated geologic structure, such as a limestone pinnacle, might not be detected by a routine drilling program (Anderson et al., 2008). Geophysical exploration is a very effective endeavour to detect even a pinnacle and map the dimension of the target bodies.

Other advantages of geotechnical geophysics are related to site accessibility, portability, non-invasiveness, and operator safety. The deployment of geophysical equipment can be made at various places, such as beneath bridges and power lines, in forested areas, at contaminated sites, in urban and rural areas, in marshy or undulated terrain, which are not appropriate for setting up drill rigs or cone penetration test (CPT) rigs. In addition, geophysical surveys are generally considered less dangerous than drilling since there are fewer risks associated with utility encounters and operations. A geophysical survey provides a yardstick for engineers to decide borehole locations judiciously, which helps in reducing the number of required boreholes for resource assessment.

1.7 GEOPHYSICAL EXPLORATION OF LITHIUM RESERVES

We know that the prosperity of any nation is measured on the availability of costly and necessary mineral resources having implications to both energy and health security. Recently, several countries like India (Geological Survey of India [GSI] assessed 5.9 million tonnes [MT] of lithium in the Salal–Haimana area of the Reasi District of Jammu and Kashmir, 5.5% of globally existing reserves [timesofindia.indiatimes.com]). Australia, Chile, China, Argentina, Zimbabwe, Portugal, Brazil, and others have been involved in the exploration of green energy resource minerals, such as lithium, one of the precious and very significant mineral resource for energy-intensive industries. Recent pegmatite-hosted spodumene (naturally occurring compound source of lithium) discoveries in Western Australia, as well as the recent discovery of lithium reserves in the state of Jammu and Kashmir of India, gave deeper insights about the geophysical exploration techniques used for mineral exploration. Due to the weak magnetic and conductive properties of lithium–caesium–tantalum (LCT) pegmatites located in Australia, geophysical surveys cannot readily distinguish between pegmatite and adjacent granite intrusions, although geophysics can be used to determine appropriate geotectonic settings (Steiner, 2019), whilst the use of different geophysical methods integrated with geochemical analysis helped in deciphering lithium resources in Australia (Phelps-Barber et al., 2022). The use of district-scale geological surface mapping identified the targets of major new spodumene-bearing pegmatite reserves at Pilgangoora in Pilbara Craton and at Earl Grey/Mt Holland, Kathleen Valley, Buldania (Anna); the Pioneer Dome North Lithium Project, Mt Ida; and the Manna in the Yilgarn Craton (Phelps-Barber et al., 2022).

Portable X-ray fluorescence (XRF) does not test for Li or Cs, and whereas in situ analysis of Sn, Ta, W, and K/Rb has proven to be a valid technique for investigation, one of the main difficulties in in situ exploration is visually identifying Li-bearing minerals (Steiner, 2019). The deposit comprises a gently dipping ($\sim 15^\circ$ NW) main body, 30–90 m thick, with hanging wall and footwall spread up to 30 m thick (Phelps-Barber et al., 2022). Spodumene is the dominant Li mineral with secondary petalite ($\text{LiAlSi}_4\text{O}_{10}$) along the western and eastern margins of the deposit (Kidman

Resources, 2019) to delineate the surface expression of the known pegmatite (Phelps-Barber et al., 2022; Kidman Resources, 2019), Mt Mann and Kathleen's Corner (Phelps-Barber et al., 2022). Pegmatites at these locations come together at a 300–400 m depth, forming a single, 35–75 m thick body which remains open with the current extent of 600–700 m down dip (Liontown Resources, 2020, 2021a). The extended depth of pegmatites up to 300 m, with a combined average thickness of 26 m, can be well deciphered using an integrated geophysical method (Liontown Resources, 2021b) for lithium exploration prior to 2017 based on comprehensive geological mapping and petrological analysis (Liontown Resources, 2017).

The exploration of Nickel (Ni) and gold (Au) during 1970–1990 has been made by identifying spodumene-bearing pegmatite swarms, as witnessed by the discovery of the Anna deposit in Holland. Further exploration, utilizing a soil sampling programme, delineated along three different trending anomalies adjacent to the Anna pegmatite deposit. Future exploration at Buldania is projected to extend the soil sampling programme for the known anomalies (Phelps-Barber et al., 2022; Liontown Resources, 2021b). The deposit is found located in the Eastern Goldfields of Yilgarn Craton. The beauty of the geophysical exploration is that it detects pegmatite hosted into the mafic and ultramafic sequences even intruded by large-scale granites inside the Mt Ida/Ularring greenstone belt. Pegmatites are spodumene-bearing, but their dimensions are undefined in the study (Red Dirt Metals, 2021). Near the Timoni gold mine, an aeromagnetic survey confirmed the presence of an adjacent granite intrusion, with associated geological mapping uncovering outcropping pegmatites at different locations (Red Dirt Metals, 2021).

Groves et al. (2022) demonstrate that globally a critical characteristic of economic spodumene pegmatite bodies is a gentle to sub-horizontal dip, which is very much corroborative with the model of emplacement in dilation zones associated with compressional and translational settings, showing multiple pulses of Li-rich melt dictated by the near-vertical orientation of minimum stress (Figure 7, Phelps-Barber et al., 2022, p. 109). Steiner's (2019) workflow for grassroots LCT pegmatite target generation follows similar pathways to the exploration for other deposit types: desktop study, regional surveys, and geological mapping, followed by targeted drilling. Through researching the exploration strategies of companies exploring for spodumene pegmatites in Western Australia, additional methods to constrain the targeting of pegmatites, such as re-assaying historical gold or base metal exploration using drill core, along with regional-scale soil sampling, have been identified. In existing gold, nickel, or base metal mining districts, re-logging and re-assaying of existing exploration drill core can be an additional tool to locate subsurface spodumene pegmatites.

1.8 INTEGRATION OF GEOPHYSICAL, GEOCHEMICAL, AND GEOLOGICAL DATASETS

Integrated geophysical methods in association with geochemical and geological data help us extract a wealth of information with better reliability in comparison to any of the single geophysical methods. For example, having used aeromagnetics to accurately locate and map a prospective contact, surface EM and electrical surveys may locate the presence of conductive unit(s) in the vicinity of the contact. Integration of geological, geophysical, and geochemical studies helped identifying new potential mineralized target areas for Pb-Zn-Au in the Rajasthan region of India (Mathur et al., 2003; Gopal et al., 2008). Resolution of the subtle structural features, along with the determination of the contact dip of the mineral bearing zones at the subsurface, can be achieved by deploying both magnetic and gravity exploration methods (Gupta and Ganguli, 2020; Bangarubabu et al., 2022). The challenge is, of course, to provide maximum information and thereby obviate excessive drilling, within the tight constraints of an exploration budget. It is very much imperative for an accomplished geophysicist to take a judicious decision for deployment-integrated geophysical methods for authenticating the target bodies prior to drilling. The usefulness of various individual techniques to the solution of common geological problems in the nickel-sulphide ore environment is shown in Table 1.2.

Beyond solely geophysical considerations, the nickel exploration philosophy draws on geophysical, geochemical, and geological datasets. For example, the termination of an electrically polarizable

TABLE 1.2

Summary of the Relative Usefulness of Geophysical Techniques to Address Common Geological Exploration Problems in the Goldfields Province of Kambalda District, Western Australia

Problem	AM	GM	Gravity	TEM	AEM	DHEM	IP	SP
Direct detection of blind nickel sulphides, massive	*	*	*	***	*	***	**	**
matrix disseminated	*	*	—	*	—	**	***	*
Massive prospective contact at the surface	**	***	**	*	*	—	**	—
Dip determination and geometry of prospective contact	**	***	***	*	*	*	*	—
Detection of stratigraphic thickening of ultramafic unit	**	***	*	*	*	—	*	—
Discrimination of nickel sulphides from sulphidic sedimentary rocks	**	**	—	**	*	*	*	—
Location of sediment-free contact	—	—	—	*	*	*	*	—
Way-up determination of ultramafic units	*	*	—	—	—	—	—	—
Discrimination of basal contact from other ultramafic–mafic contact (e.g. faulted)	*	*	*	*	*	—	—	—
Thickness and/or lateral variations in regolith cover	*	**	***	**	***	*	*	—

Source: Trench and Williams (2019).

Note:

* Possible under favourable conditions.

** Probable under favourable conditions.

*** Highly applicable—not applicable.

AM, aeromagnetic; GM, ground magnetic; TEM, transient electromagnetic; AEM, aero electromagnetic; DHEM, downhole electromagnetic; IP, induced polarization; SP, self-potential.

unit, identified geologically as likely to be an interflow sedimentary unit, at a nickel geochemical anomaly is regarded as highly prospective. The typical characteristics of a sequential exploration approach include the following:

1. Airborne magnetic data reveal the morphology of the contact regarded as prospective from independent geological and geochemical datasets.
2. Subsurface conductive structures are initially detected by ground magnetics and surface time domain electromagnetics (TEM), then different sources of sulphides/graphites in the sedimentary units along with the nickel ore lenses can be delineated based on the assessment of multiple TEM anomalies.
3. Drilling of the prospective contact, based upon the combined geological and geophysical interpretation, yields anomalous nickel values.
4. In order to detect the additional prospective mineral resources, downhole electromagnetic (DHEM) along with physical property–based loggings in drilled boreholes can be deployed to identify the geophysical anomalous bodies having intersection with host rocks.

REFERENCES

- Anderson, L.A., 1984. Self-potential investigations in the Puhimau thermal area, Kilauea Volcano, Hawaii. *SEG Expanded Abstract*, vol. 3, pp. 84, DOI: 10.1190/1.1894270.
- Anderson, N., Kansas, N.C., Dawood, R.H. and Sirles, P., 2008. Geophysical Methods Commonly Employed for Geotechnical Site Characterization. Transportation Research Board, Transportation research circular E-C130, ISSN 0097-8515.

- Apollo 17 mission, 1972. The Blue Marble, the First Full-View Photograph of the Planet, was Taken by Apollo 17 Astronauts Enroute to the Moon in 1972. NASA's Apollo Program, Payload America (CM-114), Challenger (LM-12).
- Atchuta Rao, D. and Ram Babu, H.V., 1983. Quantitative interpretation of self-potential anomalies due to two-dimensional sheet-like bodies. *Geophysics*, vol. 48, pp. 1659–1664.
- Bangarubabu, S., Kulkarni, A.V., Lakshmana, M. and Gupta, D., 2022. Regional magnetic surveys in part of Delhi Fold Belt, Rajasthan, India and implications for basement structures and mineralization. *Journal of Indian Geophysical Union*, vol. 26, no. 3, pp. 165–181.
- Biswas, U.K., Mishra, O.P. and Ghosh, D.C., 1999. An Integrated Approach for Studying the Geophysical Response over Sub-Surface Coal Fire in Raniganj Coalfield, West Bengal, India, Part I, 1998–1999. Geological Survey of India (GSI), Unpublished Report RP/CHQ/CGD/1998/009, p. 25.
- Caglar, I., 2000. Short note visual interpretation of superposed self-potential anomalies in mineral exploration. *Computational Geosciences*, vol. 26, pp. 847–852.
- Chalke, T. and McGaughey, J., 2015. Realising the benefit of integrated interpretation in minimising 3D model uncertainty. Conference abstracts: Saying goodbye to a 2D Earth conference abstracts 2015. Government of Western Australia, DMP Bookshop.
- Corry, C.E., 1981. *The Role of the Self-Potential Method in the Exploration for Molybdenite AMAX*. Golden, CO: Climax Molybdenum Co., pp. 53 (7 maps).
- Corwin, R.F., 1984. The self-potential method and its engineering applications: An overview. 54th Ann. Int. Meeting Soc. Explor. Geophys. Expanded Abstr. Society of Exploration Geophysicists, Tulsa, Session, SP1.
- Corwin, R.F., 1990. The self-potential method for environmental and engineering applications. In: Ward, S.W. (ed), *Geotechnical and Environmental Geophysics*, vol. 1. Houston, TX and Tulsa: Society of Exploration Geophysicists (SEG), pp. 127–145.
- Corwin, R.F. and Hoover, D.B., 1979. The self-potential method in geothermal exploration. *Geophysics*, vol. 44, pp. 226–245.
- Criss, R.E. and Champion, D.E., 1984. Magnetic properties of granitic rocks from the southern half of the Idaho batholith—Influences of hydrothermal alteration and implications for aeromagnetic interpretation. *Journal of Geophysical Research*, vol. 89, no. B8, pp. 7061–7076.
- Dentith, M. and Mudge, S.T., 2014. *Geophysics for the Mineral Exploration Geoscientist*. Cambridge: Cambridge University Press.
- Dickson, B.L. and Scott, K.M., 1997. Interpretation of aerial gamma ray surveys—Adding the geochemical factors. *AGSO Journal of Australian Geology and Geophysics*, vol. 17, pp. 187–200.
- Durrance, E.M., 1986. *Radioactivity in Geology, Principles and Applications*. West Sussex: Ellis Horwood Series in Geology, p. 441.
- Flood, C.L. and Aitchison, J.C. (eds), 1993. *New England Orogen, Eastern Australia*. Armidale: University of New England.
- Fullagar, P.K. and Pears, G.A., 2007. Towards geologically realistic inversion. In: Milkereit (ed), *Proceedings of Exploration 07, Fifth Decennial International Conference on Mineral Exploration, Toronto*, Corpus ID 163163020, Geology, Brisbane, Queensland, Australia.
- Fullagar, P.K., Pears, G.A. and McMonnies, B., 2008. Constrained inversion of geological surfaces—Pushing the boundaries. *The Leading Edge*, vol. 27, pp. 98–105.
- Fullagar, P.K., Pears, G.A. and Reid, J.E., 2013. Hybrid 1D/3D geologically constrained inversion of airborne TEM data. Extended Abstract, ASEG 23rd International Geophysical Conference and Exhibition, August, Melbourne, Australia.
- Fullagar, P.K., Vrbancich, J. and Pears, G.A., 2010. Geologically-constrained 1D TEM inversion. Extended Abstract, ASEG Conference, Sydney.
- Geosoft Program Package (Oasis Montaj), 1998. *Geosoft Mapping and Application System, Inc.* Toronto, ON, Canada.
- Ghosh, D.C. and Mishra, O.P., 2000. An integrated approach for studying the geophysical response over sub-surface coal fire in Raniganj coalfield, West Bengal, India, Part II, 1999–2000, RP/CHQ/CGD/1998/009, p. 28.
- Gopal, R., Gupta, D. and Radheshyam, 2008. Drill-hole induced polarisation, SP and magnetic surveys: Case studies from mineral belts of Rajasthan. *Journal of Geological Society of India*, vol. 71, p. 98–106.
- Groves, D.I., Zhang, L., Groves, I.M. and Sener, A.K., 2022. Spodumene: The key lithium mineral in giant lithium-cesium-tantalum pegmatites. *Acta Petrologica Sinica*, vol. 38, pp. 1–8.
- Gupta, D., 2006. Borehole geophysics and Mise-a-la-masse surveys geophysical prospecting for mineral exploration and mapping. *Geological Society of India*, vol. 69(1), pp. 203–204.

- Gupta, D., 2007. Mise-a-la-masse—A cost effective method of mineral exploration. *Journal of Geological Society of India*, 69.
- Gupta, D. and Ganguli, S.S., 2020. Tectonic framework revealed by gravity surveys at Varkala Beach, Kerala. *Indian Journal of Geosciences*, vol. 74, no. 3.
- Gupta, D. et al., 2016. Investigations of continued reservoir triggered seismicity at Koyna, India. In: Mukherjee, S., Misra, A.A., Calves, G. and Nemchok, M. (eds), *Tectonics of the Deccan Large Igneous Province*. London: Geol. Soc., Spec. Publ., vol. 445, DOI: 10.1144/SP445.11.
- Hanna, W.F., 1969. *Negative Aeromagnetic Anomalies over Mineralized Areas of the Boulder Batholith*. Montana: U.S. Geological Survey Professional Paper 650-D, pp. 159–167.
- Hoover, D.B., Grauch, V.J.S., Pitkin, J.A., Krohn, D. and Pierce, H.A., 1991. Getchell trend airborne geophysics—an integrated airborne geophysical study along the Getchell trend of gold deposits, North-Central Nevada. In: Raines, G.L., Lisk, R.E., Schafer, R.W. and Wilkinson, W.H. (eds), *Geology and Ore Deposits of the Great Basin*. Nevada: Geological Society of Nevada, vol. 2, pp. 739–758.
- Hope, M. and Anderson, S., 2015. Geophysical response of the Atlántida Cu-Au porphyry deposit, Chile—An undercover discovery in an old district. Extended Abstract, ASEG 24th International Geophysical Conference and Exhibition, February, Perth, Australia.
- Jaques, A.L., Wellman, P., Whitaker, A. and Wyborn, D., 1997. High resolution geophysics in modern geological mapping. *AGSO Journal of Australian Geology and Geophysics*, vol. 17, pp. 159–173.
- Jessell, M.W., Ailleres, L., de Kemp, E.A., Lindsay, M., Wellmann, F., Hillier, M., Laurent, G., Carmichael, T.L. and Martin, R.J., 2014. *Next Generation Three-Dimensional Geologic Modeling and Inversion*, *Society of Economic Geologists*. Special Publications Series [P], vol. 18, Littleton United States: Society of Economic Geologists, Inc., pp. 261–272.
- Joly, A., Porwal, A., McCuaig, T.C., Chudasama, B., Dentith, M.C. and Aitken, A.R.A., 2015. Mineral systems approach applied to GISbased 2D prospectivity modelling of geological regions: Insights from Western Australia. *Ore Geology Reviews*, vol. 71, pp. 673–702.
- Khare, N., Mishra, O.P. and Taguchi, S., 2018. Recent advances in climate science of polar region: To commemorate the contributions of late Dr. S. Z. Qasim, a pioneering doyen of the Indian polar programme. *Polar Science*, vol. 18, pp. 1–4.
- Kiberu, J., 2002. Induced polarization and resistivity measurements on a suite of near surface soil samples and their empirical relationship to selected measured engineering parameters. MSc Thesis. International Institute for Geo-information Science and Earth Observation (ITC), Enschede, The Netherlands, p. 119.
- Kidman Resources, 2019. *Quarterly Activities Report—March 2019*. Sydney: ASX, Australian Stock Exchange Announcement.
- Klein, J.D. and Sill, W.R., 1982. Electrical properties of artificial clay-bearing sandstone. *Geophysics*, vol. 47, pp. 1593.
- Kumar, A.K., Negi, S.S., Ningthoujam, L.S. and Pandey, D.K., 2022. Surface wave phase velocity variations underneath the Indian Ocean geoid low. *Tectonophysics*, DOI: 10.1016/j.tecto.2022.229688.
- Kumwenda, J. and Lackie, M., 2019. Geophysical interpretation of the geology of the Stanthorpe region using aeromagnetic, gravity and radiometric data. *Exploration Geophysics*, vol. 50, no. 6, pp. 653–666, DOI: 10.1080/08123985.2019.1659102.
- Li, Y. and Oldenburg, D.W., 1996. 3D inversion of magnetic data. *Geophysics*, vol. 61, pp. 394–408.
- Lindsay, M., Perrouty, S., Jessell, M.W. and Ailleres, L., 2013. Making the link between geological and geophysical uncertainty: Geodiversity in the Ashanti Greenstone Belt. *Geophysical Journal International [PJ]*, vol. 195, no. 2, Oxford, UK: Oxford University Press, pp. 903–922.
- Liontown Resources, 2017. *Liontown Acquires Highly Prospective Lithium Projects in WA's Eastern Goldfields*. Sydney: ASX, Australian Stock Exchange Announcement.
- Liontown Resources, 2020. *Kathleen Valley Confirmed as a World-Class Lithium Deposit as Mineral Resource Increases to 156 Mt @ 1.4% Li₂O*. Sydney: ASX, Australian Stock Exchange Announcement.
- Liontown Resources, 2021a. *Strong Progress with Kathleen Valley Definitive Feasibility Study as Ongoing Work Identifies Further Key Project Enhancements*. Sydney: ASX, Australian Stock Exchange Announcement.
- Liontown Resources, 2021b. *Kathleen Valley Confirmed as a World-Class Lithium Deposit as Resource Increases to 156 Mt @ 1.4%*. Sydney: ASX, Australian Stock Exchange Announcement.
- Macnae, J.C., 1979. Kimberlites and exploration geophysics. *Geophysics*, vol. 44, pp. 1395–1416.
- Markiewicz, R.D., Davenport, G.C. and Randall, J.A., 1984. The use of self-potential surveys in geotechnical investigations. 54th Ann. Int. Meeting, Soc. Explor. Geophys., Expanded Abstr. Society of Exploration Geophysicists, Tulsa, Session: SP6.
- Marshall, D.J. and Madden, T.R., 1959. Induced polarization, a study of its causes. *Geophysics*, vol. 24, pp. 790–816.

- Mathur, K.N., Goyal, R.S. and Gupta, D., 2003. Integration of geological, geophysical and geochemical studies to identify new potential target area for Pb-Zn in Rajasthan. (*GoldMedal*) *Journal of Geophysics*, vol. 24, no. 1, pp. 3–24.
- McGaughey, J., 2006. The common earth model: A revolution in mineral exploration data integration. In: Harris, J.R. (ed), *GIS Applications in the Earth Sciences*. Westmount, QC: Geological Association of Canada Special Publication, no. 44, pp. 567–576.
- Mishra, O.P., 1990. Interpretation of magnetotelluric data for layered earth model by the method of stripping. Dissertation Cum Thesis for Master of Science and Technology in Applied Geophysics. Indian School of Mines, Dhanbad, India, p. 25.
- Mishra, O.P., 2004. Lithospheric heterogeneities and seismotectonics of NE Japan forearc and Indian regions. D.Sc. Thesis. GRC, Ehime University, Japan, p. 223.
- Mishra, O.P., 2014. Intricacies of the Himalayan seismotectonics and seismogenesis: Need for integrated research. *Current Science*, vol. 106, no. 2, pp. 176–187.
- Mishra, O.P., 2020. Seismic microzonation study of South Asian cities and its implications to Urban risk resiliency under climate change scenario. *International Journal of Geosciences (IJG)*, vol. 11, no. 4, DOI: 10.4236/ijg.2020.114012.
- Mishra, O.P., Mandal, H.S., Singh, P., Mahato, R., Gera, S.K., Kumar, V., Ghangas, V., Sharma, B., Shekhar, S., Gusain, P., Prajapati, S.K., Tiwari, A. and Jaladi, S., 2022. Earthquake risk resilient structures and seismic microzonation of Indian cities: Strategy for safer design of structures. In: *Economic and Social Impacts of Earth Sciences*. Springer.
- Mishra, O.P., Singh, P., Ram, B., Gera, S.K., Singh, O.P., Mukherjee, K.K., Chakraborty, G.K., Chandrasekhar, S.V.N., Selinraj, A. and Som, S.K., 2020. Seismic site-specific study for seismic microzonation: A way forward for risk resiliency of vital infrastructure in Sikkim, India. *International Journal of Geosciences (IJG)*, vol. 11, no. 3, pp. 125–144, DOI: 10.4236/ijg.2020.113008.
- Mitchinson, D., Phillips, N., Perron, G., Williston, C., McMonnies, B. and Gilmore, K., 2014. District scale targeting over the Chisel and Lalor properties using multi-disciplinary criteria and weights of evidence methods: in Exploration for Deep VMS Ore Bodies—The Hudbay Lalor Case Study. Manitoba Geological Survey Symposium, November 22nd, Winnipeg.
- Monteiro Santos, F.A., 2009. Inversion of self-potential anomalies using particle swarm optimization (submitted).
- Murthy, B.V.S. and Haricharan, P., 1984. SP anomaly over double line of poles. Interpretation through Log Curves. *Proceedings of the Indian Academy of Sciences, Earth and Planetary Sciences*, pp. 437–445.
- Mustard, R., Ulrich, T., Kamenetsky, V.S. and Mernagh, T., 2006. Gold and metal enrichment in natural granitic melts during fractional crystallization. *Geological Society of America*, vol. 34, pp. 85–88.
- NCPOR-MoES, 2018. Mid-Ocean Ridges and Other Geological Features of the Indian Ocean. SCOR-Inter-Ridge Workshop: Organized jointly by Scientific Committee on Oceanic Research & InterRidge. Hosted jointly by CSIR-National Institute of Oceanography, Goa ESSO-National Centre for Polar and Ocean Research, Goa.
- Negi, S.S., Kumar, A.K., Ningthoujam, L.S. and Pandey, D.K., 2022. Mapping the mantle transition zone beneath the Indian Ocean geoid low from Ps receiver functions. *Tectonophysics*, vol. 831, p. 229330, DOI: 10.1016/j.tecto.2022.229330.
- Osiensky, I.L. and Donaldson, P.R., 1994. A modified mise-a-la-masse method for contaminant plume delineation. *Ground Water*, vol. 32, no. 3, pp. 448–457.
- Pandey, D.K., Ningthoujam, L.S., Yadav, R., Nair, N., Negi, S.S., Kumar, A.K. and Khogen, K.S., 2022. Seismic investigations around an aseismic Comorin ridge, Indian Ocean. *Journal of Geological Society of London*, vol. 179(6), pp. 615–623, jgs2021113. doi: <https://doi.org/10.1144/jgs2021-113>.
- Pant, P.R. and Gupta, D., 1998. Simple electrical logging technique for basemetal exploration. *Journal of Pure and Applied Geophysics*, vol. 152, pp. 759–772.
- Pears, G. and Chalke, T., 2019. Geological and geophysical integrated interpretation and modelling techniques. *ASEG Extended Abstracts*, Taylors and Francis, vol. 1, p. 7.
- Pears, G., Fullagar, P. and Andrews, P., 2001. 3D gravity modelling and interpretation for the 1:250,000 Boulia map sheet, Queensland. Extended Abstract, ASEG 15th Geophysical Conference and Exhibition, August, Brisbane.
- Phelps-Barber, Z., Trench, A. and Groves, D.I., 2022. Recent pegmatite-hosted spodumene discoveries in Western Australia: Insights for lithium exploration in Australia and globally. *Applied Earth Science (Transactions of the Institution of Mining and Metallurgy B)*, vol. 131, no. 2, pp. 100–113.

- Ramadan, T.M. and Sultan, A.S., 2004. Integration of remote sensing, geological and geophysical data for the identification of massive sulphide zones at Wadi Allaqi area Middle East J., Ain Shams Univ. *Earth Science Series*, vol. 18, pp. 165–74.
- Red Dirt Metals, 2021. *Pegmatites Now Drilled over 4 km of Strike*. Sydney: ASX, Australian Stock Exchange Announcement.
- Reynolds, R.L., Rosenbaum, J.G., Hudson, M.R. and Fishman, N.S., 1990. Rock magnetism, the distribution of magnetic minerals in the Earth's crust, and aeromagnetic anomalies. In: Hanna, W.F. (ed), *Geologic Applications of Modern Aeromagnetic Surveys*. Drive Reston, VA: U.S. Geological Survey Bulletin, 12201, pp. 24–45.
- Richard, S.W. and Harold, D.C.M., 1973. Radar geology-petroleum exploration technique, Eastern Panama and Northwestern Colombia. *American Association of Petroleum Geologists*, vol. 57, no. 5, pp. 825–840.
- Roy, S., 2017. Scientific drilling in Koyna region, Maharashtra. *Current Science*, vol. 112, no. 11, pp. 2181–2181.
- Roy, S., Rao, N.P., Akkiraju, V.V., Goswami, D., Sen, M., Gupta, H.K., Bansal, B.K. and Nayak, S., 2013. Granitic basement below Deccan traps unearthed by drilling in the Koyna Seismic zone, Western India. *Journal of Geological Society of India*, vol. 81, pp. 289–290.
- Schulz, R., 1985. Interpretation and depth investigation of gradient measurements in direct current geoelectric. *Geophysical Prospecting*, vol. 33, pp. 1240–1253.
- Seguin, M.K., Gupta, O.P., Nagrajan, V., Jain, A.K., Sen, N., Mahapatra, S. and Mishra, O.P., 1994. Identification and delineation of abandoned, unknown and surveyed galleries in Raniganj coalfield by geophysical methods, Exploration Technology for modern coal mines: *Coal Exploration. Seminar*, vol. VI-49–VI-70.
- Singh, A.P., Mishra, O.P., Yadav, R.B.S. and Kumar, D., 2012. A new insight into crustal heterogeneity beneath the 2001 Bhuj earthquake region of northwest India and its implications for rupture initiations. *Journal of Asian Earth Sciences*, vol. 48, pp. 31–42.
- Sirles, P.C., 2006. NCHRP Synthesis 357: Use of Geophysics for Transportation Projects. Transportation Research Board of the National Academies. Washington, DC. Available at http://onlinepubs.trb.org/onlinepubs/nchrp/nchrp_syn_357.pdf.
- Smith, R.J., 2002. *Geophysics of Iron Oxide Copper-Gold Deposits Hydrothermal Iron Oxide Copper—Gold and Related Deposits. A Global Perspective*, vol. 2. Adelaide: Porter T.M. Adelaide PGC Publishing, pp. 357–367.
- Spampinato, G.P.T., Ailleres, L., Betts, P.G. and Armit, R.J., 2015. Crustal architecture of the Thomson Orogen in Queensland inferred from potential field forward modelling. *Australian Journal of Earth Sciences [P]*, vol. 62, no. 5, Abingdon Oxon, UK: Taylor & Francis, pp. 581–603.
- Steiner, B., 2019. Tools and workflows for grassroots Li-Cs-Ta (LCT) pegmatite exploration. *Minerals*, vol. 9, p. 499, DOI: 10.3390/min9080499.
- Sternberg, B.K. and Oehler, D.Z., 1990. Induced polarization in hydrocarbon surveys: Arkoma basin case histories. In: Ward, S.H. (ed), *Induced Polarisation: Applications and Case Histories*, vol. 4. Houston, TX: Society of Exploration Geophysicists.
- Sultan, S.A., Mansour, S.A., Santos, F.M. and Helaly, A.S., 2009. Geophysical exploration for gold and associated minerals, case study: Wadi El Beida area, South Eastern Desert, Egypt. *Journal of Geophysics and Engineering*, vol. 6, no. 4, pp. 345–356.
- Sumner, J.S., 1976. *Principles of Induced Polarization for Geophysical Exploration*. Amsterdam: Elsevier, p. 277.
- Tejale, P.M., 2009. Characterization of the Airborne Anomalies Associated with Lead-Zinc Deposits at Rajpura-Dariba: its implications in Prognostication of Newer Resources in Adjoining Areas. Ph.D Thesis submitted to Department of Geology, University of Rajasthan, Jaipur (India), 218p.
- Towel, J.N., Anderson, R.G., Pelton, W.H., Olhoeft, G.R. and LaBrecque, D., 1985. Direct detection of hydrocarbon contaminants using induced polarization method SEG Meeting, pp. 145–147.
- Trench, A. and Williams, P.K., 2019. Nickel: Application of geophysics to nickel sulphide exploration in the Kambalda district, Western Australia. *ASEG Extended Abstracts*, 1994, vol. 1, pp. 169–180, DOI: 10.1071/ASEGSpec07_10.
- Vacquier, V., Holmes, C.R., Kintzinger, P.R. and Lavergne, M., 1957. Prospecting for groundwater by induced electrical polarization. *Geophysics*, vol. 23, pp. 660–687.
- Wilford, J., Worrall, L. and Minty, B., 2009. Radiometric map of australia provides new insights into uranium prospectivity. *Geology, Environmental Science, Symonston ACT*, p. 4.
- Wright, P.M., 1981. Gravity and magnetic methods in mineral exploration. In: Skinner, B.J. (ed) *Economic Geology, 75th Anniversary Volume*. McLean, VA: Economic Geology Publishing, pp. 829–839.
- Yungul, S., 1950. Interpretation of spontaneous polarization anomalies caused by spheroidal ore bodies. *Geophysics*, vol. 15, pp. 237–246.

2 Environmental and Engineering Geophysics

2.1 INTRODUCTION

Applied geophysics (local exploration) deals with geological-geophysical exploration, engineering geophysics (geotechniques), mining geophysics, hydrogeophysics, and environmental geophysics. *General geophysics* (regional exploration) deals with the Earth's physics and global environmental geophysics. The targets of engineering and environmental investigations are situated at shallow depths. Near-surface geophysical methods are adapted from exploration geophysics. A closely spaced grid of observation points is necessary for accurate localization and characterization. We attempt to combine various geophysical methods for the enhancement of the reliability of interpretation based on the analyses of integrated geophysical data. The cost of geophysical surveys depends on the selected technique, terrain conditions, dimension of the survey site, number of survey stations, required accuracy, penetration of depth, and interpretation technique. Moreover, we generally try to adopt relatively low-cost geophysical methods, but in order to address ambiguity of interpretation, drilling at sites is often necessary to confirm the results.

Geophysical instruments are designed to map spatial variations in the physical properties of the Earth (Table 2.1). A *gravimeter*, for example, is designed to measure spatial variations in the strength of the Earth's gravitational field (Table 2.1). Most surface-based geophysical instruments suffer from their inability to resolve relatively small-scale geophysical anomalies that are indeed potentially significant, as such anomalies are generated due to variations in the physical property contrasts between causative and host materials of the subsurface. With systematic measurements of such geophysical anomalies by deploying suitable geophysical methods, geophysicists become enabled to interpret these measured variations as a wealth of data for generating geologic models of varying resolutions, as shown in Table 2.1. Nonetheless, the environmental and engineering problems require precise estimates of geophysical anomalies at shallow depths or near-surface layers, but the task of acquiring noise-free data is very challenging, since the subsurface Earth is very heterogeneous at shallow depths and near the surface. In order to address such herculean task of enhancing the high signal-to-noise ratio, geophysicists have to take special precautions for both field and geophysical instrument parameterization.

Basically, most geophysical instruments are capable of imaging small targets at shallow depths, but only large targets at greater depths. Geophysical and geotechnical surveys need to be designed in such a way that targeted depth, either shallow or deep, can get deciphered without reference to the size of the target. It is important to note that those targets with appreciable variability in the physical property contrasts are decipherable by subsurface imaging (controlled-source geophysical tomography), even located at the deeper layers, by using geophysical and geotechnical methods, in contrast to how targets located at any depth cannot be detected and delineated by subsurface imaging if the physical property contrasts do not differ adequately from those of the associated strata beneath the Earth. In a nutshell, the nature and extent of homogeneity of the subsurface strata gets enhanced with depth that reduces the physical property contrasts.

TABLE 2.1**Summary of the 12 Commonly Used Geophysical Surveying Methods for Geotechnical Investigations**

Geophysical Method	Measured Parameter(s)	Physical Property or Properties	Physical Property Model (Geotechnical Application)	Typical Site Model (Geotechnical Applications)
Shallow seismic refraction	Travel times of refracted seismic energy (<i>P</i> or <i>S</i> wave)	Acoustic velocity (function of elastic moduli and density)	Acoustic velocity–depth model, often with interpreted layer boundaries	Geologic profile
Shallow seismic reflection	Travel times and amplitudes of reflected seismic energy (<i>P</i> or <i>S</i> wave)	Density and acoustic velocity (<i>acoustic velocity</i> is a function of elastic moduli and density)	Acoustic velocity–depth model, often with interpreted layer boundaries	Geologic profile
Cross-hole seismic tomography	Travel times and amplitudes of seismic energy (<i>P</i> or <i>S</i> wave)	Density and acoustic velocity (<i>acoustic velocity</i> is a function of elastic moduli and density)	Model depicting spatial variations in acoustic velocity	Geologic profile
Multichannel analyses of surface waves (MASW)	Travel times of surface wave energy generated using an active source (e.g. sledgehammer)	Acoustic velocity (function of elastic moduli and density)	Acoustic (shear wave) velocity–depth model, often with interpreted layer boundaries	Geologic profile
Refraction microtremor (ReMi)	Travel times of passive surface wave energy	Acoustic velocity (function of elastic moduli and density)	Acoustic (shear wave) velocity–depth model, often with interpreted layer boundaries	Geologic profile
Ground-penetrating radar (GPR)	Travel times and amplitudes of reflected pulsed EM energy	Dielectric constant, magnetic permeability, conductivity, and EM velocity	EM velocity/depth model with interpreted layer boundaries	Geologic profile
Electromagnetics (EM)	Response to natural-induced EM energy	Electrical conductivity and inductivity	Conductivity-depth model, often with interpreted layer boundaries	Geologic-hydrologic profile
Electrical resistivity	Potential differences in response to induced current	Electrical resistivity	Resistivity-depth model, often with interpreted layer boundaries	Geologic-hydrologic profile
Induced polarization (IP)	Polarization voltages or frequency-dependent ground resistance	Electrical capacitivy	Capacitivity-depth model	Model depicting spatial variations in clay content (or metallic mineralization)

(Continued)

TABLE 2.1 (Continued)
Summary of the 12 Commonly Used Geophysical Surveying Methods for Geotechnical Investigations

Geophysical Method	Measured Parameter(s)	Physical Property or Properties	Physical Property Model (Geotechnical Application)	Typical Site Model (Geotechnical Applications)
Self-potential (SP)	Natural electrical potential differences	Natural electric potentials	Model depicting spatial variations in natural electric potential of the subsurface	Hydrologic model (seepage through dam, levee, or fractured bedrock, etc.)
Magnetics	Spatial variations in the strength of the geomagnetic field	Magnetic susceptibility and remnant magnetization	Model depicting spatial variations in magnetic susceptibility of subsurface	Geologic profile or map (location of faults, variable depth to bedrock, etc.)
Gravity	Spatial variations in the strength of the gravitational field of the Earth	Bulk density	Model depicting spatial variations in the density of the subsurface, often with interpreted layer boundaries	Geologic profile or map (location of voids, variable depth to bedrock, etc.)

Source: Anderson (2006).

2.2 GEOPHYSICAL METHODS

Several approaches are adopted for addressing environmental and engineering problems using different geophysical and geotechnical methods that require different sets of parameters, as described in what follows.

1. Non-invasive, non-destructive methods (ground geophysical surveys)
2. On-site exploration work, in situ measurements (geophysical sounding in penetration holes), continuous information
3. Geotechnical and site-specific surveys
4. Time-lapse measurements (monitoring surveys)
5. Solution of forward/inverse problem using acquired geophysical data

2.2.1 GEOMETRICAL PARAMETERS

Structural elements, layer thickness, depth, dip, strike, azimuth, tectonics, volume, structure of establishments and construction works, 1-D, 2-D, 3-D, and mixed models and structures.

2.2.2 SPATIAL DISTRIBUTION OF PETROPHYSICAL/GEOPHYSICAL PARAMETERS

A set of petrophysical and geophysical parameters is estimated to understand the nature of the sub-surface formations and their latest state of disposition in the subsurface. Among those parameters are mineral composition, petrophysical properties (porosity, water saturation, hydraulic conductivity, etc.), degree of cracking and weathering, watertightness, contamination, and radiological parameters.

2.3 SHALLOW GEOPHYSICAL INVESTIGATION

Targets of engineering and environmental investigations are situated at shallow depths of few tens of meters. Near-surface geophysical methods have been adapted from exploration geophysics. A closely spaced grid of observation points is necessary for accurate localization and characterization. As mentioned earlier, ambiguity of interpretation and drilling is often necessary to confirm the results.

2.4 ENVIRONMENTAL PROBLEMS

- Location and characterization of near-surface geological structures (pore space, faults, fissures, shear zones, lithologic variation)
- Characterization of aquifers, groundwater protection from contamination, salinity of underground water (fresh- and saltwater contact)
- Landfill characterization, delineation of the margins of buried waste dumps, soil contamination, tracing seepage movement
- Identification of new potential sites for the safe disposal of nuclear and chemical waste
- Landslides and ground subsidence (e.g. hydrocarbon exploitation)
- Archaeological site delineation
- Evaluation of earthquake hazards
- Mining problems and safety (detecting tectonic disturbances and fault zones, water inrush, thickness of impervious layers)
- Environmental hygiene, radioactivity surveys for indoor radon risk and groundwater contamination, radiation from industrial and waste dumps, delineation of radioactive fallouts, detection of fracture zones (earthquake prediction)

2.5 CIVIL ENGINEERING PROBLEMS

- Testing of foundations (depth and composition of bedrock; physical properties of rocks in dams, tunnels, bridges, shafts)

- Survey of establishments and construction works (railway, highway, subway)
- Geotechnical problems (soil properties, elastic parameters, compaction)
- Blast planning and analysis (estimation of blast loading on a specific structure, modelling, and simulation; risk zones)
- Location of water (water supply, drainage problems, foundation and transport engineering)
- Location of older underground excavations (detection of abandoned mineshafts, pipelines, metallic objects)

2.6 ENVIRONMENTAL APPLICATIONS

Survey Technique	Area of Application					
	Geological Structures/Barriers	Hydraulically Active Structures	Landfills/Waste Dumps	Contamination Plumes	Landslides/ Ground Sinking	Pre-Excavation Archaeology
Gravity	+	0	0	–	0	0
Magnetic ^a	+	–	+	–	–	+
Self-potential	0	+	0	0	+	–
Resistivity + IP	+	+	+	+	+	+
Electromagnetic ^a	0	+	+	+	–	0
Ground radar	0	0	+	0	0	+
Radioactive ^a	0	0	0	0	0	–
Seismic refraction	+	0	0	–	0	0
Seismic reflection	+	0	–	–	–	–
Geothermal ^a	0	+	0	–	0	–

Note: + Applicable.

⁰ Limited applicability.

– Not applicable.

^a Technique applicable to both ground and airborne surveying.

2.7 ENGINEERING APPLICATIONS

Survey Technique	Area of Application					
	Geological Structures/Barriers	Hydraulically Active Structures	Landfills/Waste Dumps	Contamination Plumes	Landslides/ Ground Sinking	Pre-Excavation Archaeology
Gravity	+	–	–	+	–	–
Magnetic ^a	+	–	–	+	–	+
Self-potential	–	–	+	–	–	0
Resistivity + IP	+	–	+	+	+	0
Electromagnetic ^a	0	–	+	0	+	+
Ground radar	+	0	+	+	+	0
Radioactive ^a	–	–	0	–	–	–
Seismic refraction	+	+	0	0	+	–
Seismic reflection	+	+	0	0	0	–

Note: + Applicable.

⁰ Limited applicability.

– Not applicable.

^a Technique applicable to both ground and airborne surveying.

2.8 GEOPHYSICAL FIELDWORK

The acquisition of data is fundamental to geophysical investigations, and ‘real’ data is only acquired in the field (University of Miskolc, Szechenyi, 2020). Without real data, no true practical conclusions can be made about a particular requirement. The equipment for ground surface surveying is the major part of geophysical fieldwork (Figure 2.1). Thus, in the geophysical field surveys, we acquire data from the field and then assimilate a plausible model as accurate as possible with reference to the true earth model, which is the best example of geophysical inversion method.

2.9 HYDROLOGIC APPLICATIONS OF GEOPHYSICAL METHODS

A wide variety of geophysical techniques is available. Many books provide a description of the underlying physics and applications for the various methods (Telford et al., 1990). The art of the geophysicist is to pick out a particular geophysical method that is most suitable, given its sensitivity for the property that needs to be mapped, the scale that has to be represented, and the environmental noise conditions that might interfere with the geophysical survey. Here we have described some major applications of geophysics to map hydrologic variables.

2.9.1 VADOSE ZONE

The vadose zone, commonly referred to as an unsaturated zone, plays an important role associated with environmental issues, such as soil and groundwater contamination, ground stability, and flood control. Monitoring the spatial and temporal variability of moisture content and the freezing/thawing of the subsurface can yield valuable information regarding these environmental concerns. Geoelectric methods have immense potential to distinguish several contrasting bodies because of electrical resistivity and conductivity contrasts. It finds application in deciphering and distinguishing the frozen soil from the unfrozen part of the subsurface (French and Binley, 2004). Ground-penetrating radar (GPR) found its applications in permafrost areas, even helping detect buried faunas, animals, and humans during snow avalanches. Applications of GPR successfully detect



FIGURE 2.1 The ground geophysical survey equipment.

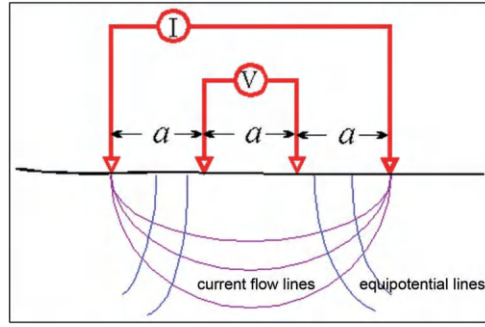


FIGURE 2.2 Measurement setup for electric resistivity tomography (ERT): (a) the spacing between the electrodes (red arrows), where I indicates an electrical current (ampere) and V indicates the potential difference (volt) that is measured during a survey.

subsurface water flushing zones, along with the detection of the localized but hidden mine fire zones in coal mine areas of India, because of appreciable contrasts in the dielectric values between the two strata (Seguin et al., 1994).

To monitor moisture content θ , two groups of geophysical techniques are currently employed. The first group of geophysical techniques comprises radar and geoelectric methods based on relating the electric permittivity and electrical resistivity of the subsurface with its water content. These geophysical techniques do not measure electric permittivity or resistivity directly but estimate these values based on the measured data. For radar methods such as time domain reflectometry (TDR) (Michot et al., 2003) and ground-penetrating radar (GPR) (Seguin et al., 1994; Knight, 2001), measured data comprise recorded electromagnetic wave velocities, which are obtained by transmitting an electromagnetic wave, after which their refracted and reflected waves are recorded. Figure 2.2 represents the setup of a geoelectric survey using electric resistivity tomography (ERT), in which electrodes are placed in the subsurface, after which potential differences (V) are measured by applying an electrical current (I) for different electrode combinations with a spacing (Naskar and Das, 2009; Herckenrath, 2012).

Based on the potential differences (voltages) and the used electrode configuration (Wenner, Schlumberger), an apparent resistivity can be calculated. Apparent resistivities can be computed from subsurface resistivity distributions using the forward model for direct current (DC) electric surveying. Matching measured and simulated apparent resistivities results in estimated subsurface resistivity distributions. For radar methods, the estimated electrical permittivity can be related to soil moisture content by an empirical petrophysical relationship, called the Topp equation (Topp et al., 1980):

$$\theta = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} \varepsilon_r^2 + 4.3 \times 10^{-6} \varepsilon_r^3 \quad (2.1)$$

where ε_r is the electrical permittivity of the subsurface normalized over the permittivity of free space, also known as the relative permittivity, or dielectric constant. Examples of studies using this approach to estimate θ can be found in Kowalsky et al. (2005), Lambot et al. (2009), and Huisman et al. (2003). A more general overview of environmental applications for GPR can be found in Knight (2001). When using resistivity methods, often, Archie's law is used to estimate the moisture content, given by Looms et al. (2008) as:

$$\theta^n = r_w \frac{\theta_s^{n-m}}{r_b} \quad (2.2)$$

where r_w indicates the electrical resistivity of water, and n and m are shape factors that are soil-specific (Looms et al., 2008). r_b represents the resistivity of the bulk material. Note that r_b is the estimated parameter using a geoelectric method. Earlier, eq. 2.2 is applied for estimating moisture content; a reasonable estimate is required for r_w . The first group of techniques is not very attractive for clayey sediments, as radar methods suffer from a limited depth of penetration due to dielectric dispersion in sediments with low electrical resistivity (Knight, 2001). For geoelectric methods, Archie's law does not apply anymore, as this empirical law does not include surface conductivity through the bulk material itself (Lesmes and Friedman, 2005). This additional term is complicated to characterize and makes the estimation of soil moisture content less reliable. The second group of methods does not require a relationship like Archie's law or the Topp equation but directly relates the measured geophysical signal with the moisture content of the subsurface in the geophysical forward model. Two of these methods are nuclear magnetic resonance (NMR), or magnetic resonance sounding (MRS) (Legchenko and Valla, 2002), and time-lapse gravity (Montgomery, 1971).

With relative gravimeters, the vertical component of the gravitational acceleration is measured. This reveals spatial differences in density which can be due to ore deposits, buried palaeochannels, and changing depth to bedrock (Carmichael and Henry, 1977; Zawila et al., 1997). To monitor changes in mass for monitoring natural gas extraction (van Gelderen et al., 1999) and quantifying changes in water storage (Christensen et al., 2011), time-lapse relative gravity (TL-RG) is found to be the best tool for the purpose. MRS is commonly known through its application in hospitals, where magnetic resonance imaging (MRI) is used for imaging and diagnosis. With MRS, the spins of the hydrogen protons of water molecules in the subsurface are excited with an external magnetic field, applied through a transmitter loop on the ground surface. After the external magnetic field is switched off, the spins of the hydrogen protons return to their original state, generating a new magnetic field (Legchenko et al., 2002) whose magnitude is measured by a receiver on the ground (Herckenrath, 2012). Legchenko et al. (2002) show the three stages of MRS sounding. It is observed that the received MRS signal oscillates exponentially with decreasing amplitude. Typically, two properties are extracted from this signal, which are the initial amplitude of this signal, depending on soil moisture content, and an exponential decrease rate of the amplitude (relaxation constant), which correlates with the pore characteristics of the subsurface and can be used to estimate hydraulic conductivity (Mohnke and Yaramanci, 2008; Vouillamoz et al., 2008).

2.9.2 LANDFILLS AND CONTAMINANT TRANSPORT

Milosevic et al. (2011) and Lorah et al. (2009) studied low-elevation areas, such as pits and wetlands typically used for waste deposition. They showed that waste deposition at these old landfills often lacked adequate control and documentation, such that the boundaries of the landfill and the type of landfill material are completely unknown. Some of these landfills pose a significant environmental threat in polluting groundwater and surface water (Christensen et al., 2011; Lorah et al., 2009). Such landfills usually contain household, demolition, and chemical waste, where the main impact on surrounding water bodies is associated with inorganic micro-components (chloride, sodium, ammonium), dissolved organic carbon (DOC), and several different xenobiotic organic compounds (Bjerg et al., 2011; Kjeldsen et al., 2002). The heterogeneous nature of an old landfill causes high spatial variability of the leachate compounds, and a large amount of work is required to accurately delineate the landfill and detect leachate plumes.

Meju (2000) lists geoelectric and electromagnetic methods as the most popular geophysical techniques to characterize landfills due to their ability to detect changes in electrical resistivity, which correlates with moisture content and chemical composition of the pore water, and the relatively low cost to perform such surveys. Due to the presence of saline fluids in the landfill leachate, which is a good electrical conductor, it is possible to delineate the landfill and locate a contaminant plume by employing these geophysical techniques (Naudet et al., 2004; Chambers et al., 2004). One major limitation of using electrical resistivity in landfill surveys is the fact that several factors influence

the electrical resistivity of the subsurface, which makes it difficult to differentiate between one another. For example, clay and saline fluids both have a small electrical resistivity.

Induced polarization (IP) is considered an upcoming technique for delineating landfills and detecting solute plumes, which can be performed in combination with a geoelectric survey (Dahlin et al., 2002; Sogade et al., 2006). IP is based on the fact that the subsurface is able to act as an electric capacitor and store electric charge. The same measurement setup as described by Herckenrath (2012) can be used, but instead of measuring the potential differences when an electrical current is injected, the decay of these potential differences is measured in time after terminating the applied electrical current. The decay rate of the potential difference can then be related to the chargeability of the subsurface. The signal that is retrieved with IP is mainly the result of the local redistribution of ionic charge in the electric double layer at the mineral–fluid interface (Slater, 2007). Typically, the observed IP response is fitted by an empirical relationship named the Cole–Cole model (Pelton et al., 1978), in which parameters (chargeability, electrical resistivity, relaxation time, and shape parameter) can be correlated with hydraulic conductivity and the presence of contaminant plumes. The Cole–Cole model, however, does not provide a mechanistic understanding of the retrieved IP signal. A physical model is still lacking to explain the IP signature of contaminant plumes. Vaudelet et al. (2011), along with Revil and Florsch (2010) and Leroy and Revil (2009), developed a physical model that is used during investigations.

2.9.2.1 Field Sites and Data Collection: Risby Landfill

A case study of engineering geophysics is well documented for the Risby landfill, located 20 km west of Copenhagen, Denmark, to identify the dominant processes affecting the hydrogeological and geochemical interaction between the landfill, the regional aquifer, and the Nybolle stream (Herckenrath, 2012). For this purpose, boreholes have been drilled and levelled, together with the collection of indirect information using several geophysical methods. In Herckenrath (2012), a map is provided with the data that was used in the paper, comprising groundwater-level measurements, seepage measurements in the Nybolle stream, borehole information, and an ERT profile.

A detailed historical overview of the Risby landfill was provided by Thomsen et al. (2004). The geological setting of the Risby landfill (Gazoty et al., 2011) comprises pre-Quaternary limestone bedrock overlain by Quaternary glacial deposits. The pre-Quaternary limestone surface is located between –10 and +5 mamsl, corresponding to 20–30 m below the natural terrain surface. The Quaternary glacial deposits mainly consist of clay till, but intercalated sand lenses and sand layers are common. The sandy deposits range in thickness from a few centimetres to several meters. Additionally, water samples, ecological data, and other geophysical data were collected in order to quantify contaminant fluxes from the landfill towards the stream and groundwater (Milosevic et al., 2011; Thomsen et al., 2004) and identify which properties of a typical Danish landfill can be mapped using state-of-the-art geophysical techniques such as magnetic methods, induced polarization, and geoelectric methods (Gazoty et al., 2011).

2.9.3 REGIONAL GEOLOGY AND COASTAL REGIONS

To characterize regional geological properties and human structures, seismic and electromagnetic methods are often employed. With seismic methods, a seismic wave is generated by an explosion or vibrator, after which the wave is reflected and refracted at geological interfaces of materials with different seismic velocities. These reflected and refracted waves are recorded by receivers, called geophones or hydrophones (offshore applications), to obtain a seismogram. Seismic velocity depends on the density of the rock, which makes the method attractive to determine the thickness of unconsolidated sediments overlying bedrock (Miller et al., 1989). Electromagnetic surveys can be performed using ground-based and airborne instruments. In the time domain electromagnetic (TDEM) setup (figure is given in Chapter 8), a square transmitter loop is used to generate an electrical current, which is switched off to generate a magnetic field in the subsurface, whose strength

decreases after the electrical current is turned off. A receiver coil, placed in the middle of or outside the transmitter loop, is used to record this decrease in magnetic field strength, which can be translated to a series of apparent resistivities which, in turn, are inverted to obtain the subsurface resistivity distribution.

2.9.4 GEOPHYSICAL METHODS: HYDROGEOLOGICAL CASE STUDIES

Three case studies illustrate both the potential and limitations of geophysics-based site characterization and monitoring. The case studies include examples of imaging hydrogeological structures that control groundwater flow and transport, direct imaging of contaminant plumes, and time-lapse monitoring of amendment delivery. It is observed that microbial processes associated with the attenuation of groundwater contaminants result in geophysical signatures. For example, the microbial-facilitated precipitation of metallic minerals (iron oxides and iron sulphides) can result in large changes in electrical and magnetic signatures (Williams et al., 2009).

2.9.4.1 Hydrogeological Characterization and Controls on Contaminant Transport at the Hanford Facility, Richland, WA

A case study of hydrological characterization and controls on the contaminant transport at the Hanford site located in Richland, Washington, USA, is presented, which is located on the west bank of the Columbia River at the south end of the Hanford Site. Underground storage of spent nuclear waste from 1943 to 1975 in the area caused uranium contamination in the groundwater. This contaminated groundwater discharges to the Columbia River through a zone of surface water–groundwater interaction beneath and adjacent to the river. It is reported that lithologic variability interacts with river stage variations to regulate the exchange between contaminated groundwater and river water, which is a major scientific challenge at this site. It is also found that the interface between the permeable Hanford Formation and the relatively impermeable Ringold Formation is a critical hydrogeological contact, controlling the vertical flow and transport of contaminated groundwater into the Columbia River (Slater et al., 2010). The geological formation so-called the Ringold Formation is mainly divided into gravel-dominated (upper) and mud-dominated (lower) units. The overlying Hanford Formation is primarily flood deposits with permeable palaeochannel that may serve as preferential flow paths for contaminated groundwater, locally facilitating the mixing of contaminated groundwater with the river water. These palaeochannels may be locally incised below the regional Hanford–Ringold contact.

A 2-D electrical resistivity imaging at the site illustrates how geophysical methods aid in understanding contaminant movement (Slater, 2022). The role of the geophysical survey is clearly reflected when the surface-acquired resistivity images captured variations in the depth to the Hanford–Ringold contact over 400–500 m along each line (Slater, 2022), which was further confirmed by direct borehole observations at two different locations (Slater, 2022). Resistivity imaging is a very versatile methodology that provided very strong evidence of a palaeochannel incised into the Ringold Formation (Slater, 2022). It is also reported that this palaeochannel likely controls much of the discharge of uranium to the Columbia River. It is interpreted that the place where the palaeochannel directs groundwater discharge into the river found a good correlation with a zone of enhanced groundwater–surface water exchange estimated from temperature monitoring on the riverbed.

2.9.4.2 High-Resolution Characterization of a Fractured Rock Aquifer at the Former Naval Air Warfare Center (NAWC), New Jersey

The case history of high-resolution characterization of the fractured rock aquifer at the former Naval Air Warfare Centre, New Jersey, is well documented, in which the role of geophysical methods has been enumerated in assessing fractures and cracked volume of the basement rock. The former Naval

Air Warfare Center (NAWC), located in West Trenton, NJ, tested jet engines from 1955 to 1998 (Slater, 2022). The fractured bedrock aquifer at the site was contaminated with the chlorinated solvent trichloroethylene (TCE) during operations. This study studied contaminated, unweathered mudstone units, where the dominant flow pathways are through a series of cross-cutting faults and discrete fracture zones associated with carbon-rich intervals (Slater, 2022).

The advanced geophysical method of high-resolution, cross-borehole resistivity tomography imaging has immense potential to identify and delineate unweathered rock using a series of boreholes (Slater, 2022). The 3-D electrical resistivity tomography (ERT) image, shown as oblique slices, reveals a subsurface structure of alternating conductive and resistive layers that represent the strike (orientation) and dip (angle from horizontal) of alternating massive and laminated mudstone units. Interpretation is based on the fact that lighter units are associated with less-conductive massive units, and darker units correlate with more-conductive, laminated units. Fracture zone intersection depths align well with layered contacts (Slater, 2022). Identification of the fractured zones is made on a thumb rule for which fractured zones are formed along the bedding plane; the contacts between units are also visible. It is seen that the largest high-electrical-conductivity features corroborate with the most hydraulically connected units. In a nutshell, the resistivity data provides a high-resolution view of the structure of the rock mass between the boreholes. However, these images were generated using state-of-the-art 3-D imaging methods that are not accessible to most geophysical contractors (Slater, 2022).

2.9.4.3 Tracking Amendment Delivery at the Former Brandywine MD Defense Reutilization and Marketing Office (DRMO)

Geophysical tools can be deployed for time-lapse monitoring of subsurface processes, including contaminant migration and amendment delivery at different site conditions (Johnson et al., 2015a, 2015b). It is observed that changes in geophysical properties with time eliminate much of the uncertainty in data interpretation, for which inverse geophysical methods can be designed to directly image changes in geophysical properties over time in order to highlight fluid movement within the subsurface (Johnson et al., 2015a, 2015b).

The former Brandywine, MD, Defense Reutilization and Marketing Office (DRMO) has an area of an 8 ac site used to store waste and excess government material generated by several US Navy and US Air Force (USAF) installations from 1943 to 1987. It is worth mentioning that geophysical imaging is conducted to validate and demonstrate the use of autonomous time-lapse ERT as an effective amendment emplacement monitoring tool (ESTCP Project ER-200717). Time-lapse three-dimensional (3-D) imaging datasets were collected before and after amendment emplacement for approximately 2.5 years, resulting in a comprehensive ‘four-dimensional’ (4-D) dataset consisting of nearly 800 images (Slater, 2022). These datasets demonstrated the capability of ERT to effectively monitor the emplacement and migration of carbon substrate and pH buffer in 3-D following injection. During the first year, the amendment was shown to slowly sink down to a lower confining unit and spread due to groundwater flow. It is interesting to note that a large increase in subsurface conductivity within the treatment zone has been recorded in the second and subsequent years, which explained the fact that due to enhanced microbial activity, such drastic changes happened (Johnson et al., 2015a, 2015b; Slater, 2022). This increase may be the result of microbial-induced iron sulphide precipitation.

2.9.4.4 The Evaluation of Hydrogeological Condition: A Case Study on Jessore Sadar, Bangladesh

Ahmed et al. (2019) investigated the area, and it constitutes a very important case study on Jessore Sadar, Bangladesh, which is well-communicated with Dhaka and nearby districts through the connectivity to highways. It is about 177 km south-west of the capital city, Dhaka. Geomorphological characterization and delineation of the subsurface structures have been made for the study area

Jessore Sadar Upazila of Jessore District, and it is reported that the area consists of Pleistocene Madhupur clay and is underlain by fluvial-deltaic sediments as the source of the major aquifer system in the area. The shallow aquifer of the study area is severely contaminated by arsenic. More than one-third of the nation's population receives their drinking water from the public water system, which uses groundwater as their source. So it is a crucial need to provide safe drinking water. The present study has concentrated on the delineation of the hydrogeological condition for sustainable groundwater development in the study area on the basis of borehole data collected from the Department of Public Health and Engineering (DPHE) (Ahmed et al., 2019).

In order to characterize the various attributes of aquifers in an area, comprehensive geophysical bore log data provide significant information. In the study area of Jessore Sadar, 12 boreholes are available. The aquifer systems of the study area are shown in the model, which is prepared by GIS software. Topsoil is the topmost horizon of the stratigraphic sequence of the study area. This topsoil is mainly composed of silt-sized particles having a portion of sand and clay particles. The 12 boreholes confirm the presence of this unit, though with some variation from bore to bore. Borehole data suggests a variation of thickness from 2 m to 9 m for this unit. Most of the borehole data shows a thickness of around 6 m. Aquitard is the other unit of floodplain deposits of the Ganges River. This is one of the most important connecting units. Data available from boreholes shows that this aquitard unit is present in all boreholes. The borehole data show its continuity throughout the area (Ahmed et al., 2019). However, the depth and thickness are not uniform all over the area. Borehole data informs about this unit having a thickness varying from 6 m to 40 m.

The 3-D block diagram shows the lithological distribution in lateral and vertical directions within the whole study area, reflecting subsurface lithological variation that indicates changes in the depositional environment with time and space (Ahmed et al., 2019). The 3-D block diagram shows that the fine sand and medium sand layers are thicker than the clay and very fine sand layers. On the other hand, a multi-log block diagram study reported the lithology of individual borehole points (Ahmed et al., 2019), which showed a vertical and lateral variation of the rock types in the study area. The upper topsoil and the immediately underlying clay are continuous throughout the area, with varying depth and thickness. The lower sand unit occurs at greater depth in boreholes.

The Environmental Protection Agency (EPA) set an arsenic maximum contaminant level for public water supplies at 0.01 mg/L (milligram per litre). The arsenic distribution contour map of the study area based on electrical resistivity and laboratory measurements of the electron count showed the increase of arsenic contaminant towards the north-western part of the study area because of apt geological formation, whilst the south-eastern part of the survey area is comparatively free from the arsenic contaminant in Bangladesh (Ahmed et al., 2019). The Environmental Protection Agency (EPA) set a chlorine maximum contaminant level for public water supplies at 10 mg/L. The chlorine distribution contour map of the study area shows that the chlorine contaminant increases towards the south-eastern middle part of the study area, whilst the south-eastern part of the survey area has comparatively lower chlorine contaminant (Ahmed et al., 2019). Similar studies have been carried out by the Geological Survey of India for the Kolkata metropolitan city using integrated geophysical surveys (Choudhury et al., 1997).

The Environmental Protection Agency (EPA) set a maximum contaminant level for iron for public water supplies at 0.3 mg/L. The iron distribution contour map of the study area (Figure 2.3) shows that the iron contaminant increases towards the south-western part of the study area, and the north-eastern part of the survey area is with comparatively lower contaminant level. The study area of Jessore Sadar Upazila lies within the Ganges River floodplain region and consists of Pleistocene Madhupur clay and underlying Plio-Pleistocene fluvial deltaic sediments which form the major aquifer system of the area and consist of shallow and deep aquifers (Ahmed et al., 2019).

An interpretation of borehole data shows the existence of three layers in the study area. The units are topsoil; clay, or silty clay, to silt, which is also termed as *aquitard*; and sand, respectively. The silty-clay-to-silt unit, which acts as a shallow aquifer, shows high arsenic and iron concentrations. The deep aquifer of fine to medium sand is a potential source for groundwater development as it is

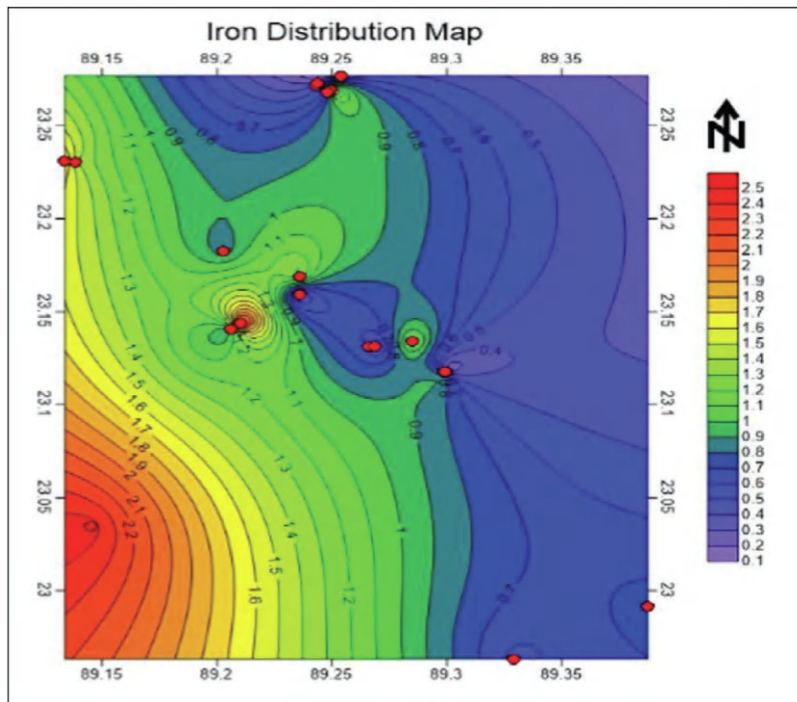


FIGURE 2.3 Iron distribution contour map for the areas of Bangladesh.

arsenic-free and as the overlying clay or silty-clay-to-silt unit seems to be continuous everywhere in the area and can act as a barrier to the vertical percolation of arsenic-contaminated water from the shallow aquifer to the deep aquifer.

2.10 DETERMINATION OF BEARING CAPACITY OF GRAVEL BOULDER DEPOSIT FROM FOOTING LOAD TEST

Geophysical surveys play a very important role in estimating the bearing capacity of gravel boulder deposits. The very successful case history of the Jammu-Udhampur-Katra-Quazigund-Baramulla railway line has been documented as the biggest project in the construction of a mountain railway since independence for the Jammu to the Baramulla, with the distance of the new rail line of about 345 km. It passes through the young Himalayas, tectonic thrusts, and faults assessed by seismological and geotechnical studies using tools of engineering geophysics. The proposed 950 m long Sangaldan station yard is located between 92.360 km and 93.310 km and is part of the USBRL project. It involves cutting from 17 m to 55 m in height. Geotechnical investigation reveals that rock is not available up to the formation level of the station yard (Dev et al., 2012).

Assessment of the bearing capacity of the soil or gravel boulder strata is one of the more challenging tasks because of the involvement of several components that make the task very difficult for the field arrangements of testing. Moreover, the load displacement of gravel boulder deposits is controlled by several factors, such as type of soil, compaction/interlocking of soil molecules, and more importantly, size of the boulders beneath the footing. The common test applied to resolve such a problem is the footing load test, which deals with the subsurface investigation in relation to the design of foundations for single- and multi-storeyed buildings, overhead water tanks, piers, and abutments of bridges in gravel boulder deposits. A recent study also highlighted that the load settlement curve can also be utilized to determine the yield pressure and recommended allowable pressure

for foundation design (Dev et al., 2012). In order to get the in situ characteristics of gravel boulder deposits, we can conduct field tests as the only tool available to civil engineers. Interpretation of the in situ footing load tests conducted at the proposed colony at the Sangaldan station yard of the Udhampur-Srinagar-Baramulla Rail Link (USBRL) project provided a clear insight into the bearing capacity and the settlement of the foundation that was determined with the help of load settlement curve for the actual footing (Dev et al., 2012). Based upon the test results of two footing load tests, the ultimate bearing capacity (at design load) of the geological formation at the proposed residential colony of the Sangaldan station yard is estimated to be 13.33 t/m². However, since the failure had not occurred at the specified design load, the ultimate bearing capacity and safe bearing capacity of the formation may be higher, which may be ascertained by applying a higher load.

2.11 MARINE GEOTECHNICAL AND GEOPHYSICAL SURVEYS IN COMPLEX GEOLOGICAL FORMATIONS

Marine geotechnical and geophysical studies in complex geological formations require special attention. As mentioned earlier, geotechnical-geophysical investigations improve the reliability of the simulated model. The high speed of geotechnical investigations and the reduced cost of the investigation make it more of a favourite among engineers and policy planners. Sometimes, geophysical tools are not always capable of addressing the objective-based requirements of highway engineers. The subsurface targets of interest may be too small or deep to resolve or impossible to effectively image because its physical properties are too low compared to those of the encompassing data. Moreover, by allowing that borehole geophysical survey control is not available, geophysical interpretations may be non-reliable and inaccurate because of their inherent non-uniqueness. Next is a case study of the Western Freeway Sea Link (WFSL-Phase IIA), which is an offshore bridge project connecting two major hubs from Worli to Haji Ali, running parallel to the western coastline of Mumbai, India (Birid, 2016). Birid (2016) showed that the bridge is an extension of the existing offshore bridge from Bandra to Worli. The main bridge will be connected to the mainland using two connector bridges at the Worli end and at the Haji Ali end. It is observed that geophysical surveys deployed in water depths varying from 3 m to 12 m establish seafloor bathymetry and map the seafloor and subsurface geological features to identify man-made structures or significant debris or pipelines within the proposed bridge corridors. The survey corridor consisted of a 200 m width throughout the length of 5.5 km along the proposed bridge alignment. Side-scan sonar and single-beam echo sounder were run within this corridor at 25 m intervals. Sub-bottom profiler, multibeam echo sounder, and magnetometer were run only along the centreline of the proposed bridge alignment to confirm results as accurately as possible.

It is also documented that geotechnical investigations detect weathered and fractured compact basalt rock predominantly encountered at or just below the seabed surface along the 4 km length of the alignment. It has been shown that this analyzed layer is believed to be that of the Malabar Hill compact basaltic lava flow (Birid, 2006). Analysis reveals that the thickness of the weathered compact basalt rock varies all along the bridge alignment but ultimately grades into fresh (unweathered) compact basalt at higher depths below the seabed. However, the intertrappean layer underlying the Malabar Hill compact basaltic flow was encountered along the east direction of the main alignment along the eastern half of the Haji Ali connector. Based on lithological and stratification data, the survey corridor is broadly divided into two distinct regions, namely, (a) main alignment with the Worli end, and (2) the Haji Ali end. The variation in stratification along the main alignment and Worli connector has also been mapped (Birid, 2016). In this study, the delineation of soil strata overlying the rock head and soft, sandy clay has also been made, along with the estimate of the depth of the soil overburden at the survey site.

The weathered and fractured compact basalt was generally encountered either immediately at the seabed level or underlying the soil overburden. The degree of weathering of the compact basalt varied from completely to moderately weathered. Fresh compact basalt rock was observed

underlying the layer of weathered compact basaltic rock. Low rock quality designation (RQD) values within the fresh compact basalt at various depths are attributed to zones of intense fracturing. The surveys also deciphered the intertrappean layers, representing a combination of tuffs, breccias, and shales. These layers were encountered predominantly within a bay along the eastern half of the Haji Ali connector to a significant depth below seabed level. Spilitic basalt rock (altered basaltic rocks that contain sodic plagioclase [albite] and chlorite) is formed because of the chemical alteration of compact basalt and was encountered only at the connector. The different areas along the bridge aligned with the predominant rock types are shown in Birid (2016). The geophysical survey revealed a generally flat seabed profile. The side-scan sonar record revealed the following reflectivity (the property of reflecting light or radiation, especially reflectance, as measured independently of the thickness of a material) patterns in the present survey area: type 1, low reflectivity (silty sands/sandy silt) over most survey areas, and type 2, high dappled reflectivity (weathered rock with jagged surfaces and rock pinnacles). The dappled reflectivity occurs possibly due to the jagged surfaces of the weathered rock and the isolated rock pinnacles. No significant features, such as pipelines, cables, and other submarine facilities, were observed in the side-scan sonar records within the survey corridor.

The correlation of the sub-bottom profiler records along with the adjacent crosslines also contributed to arriving at the distribution of the following stratigraphy units. Strata A consists of fine to medium-grained silty sands/sandy silts underlain by, strata B, a weathered bedrock. The sub-bottom profiler revealed no evidence of occurrences of buried pipelines, buried objects, shallow gas, faulting, and slumps within the survey corridor. The rapid anomaly variations in the magnetometer data were attributed to the weathered bedrock, which is compact basaltic in nature.

The connectors connecting the main bridge to the land are located within the intertidal zones, with water depths varying from 0 m during low tides to a maximum of 4 m during high tides, making it practically impossible to carry out any kind of survey in this area. The study also revealed that the ground conditions at the intertidal zone were highly undulating, with exposed compact basalt at the Worli end and soil overburden at the Haji Ali end. The two JUBs (jack-up barges) were engaged to carry out the entire geotechnical investigations with the minimum required water draft of 0.8 m and 1.5 m (Birid, 2016).

The decision to carry out a geophysical survey was beneficial and provided supplementary information on seabed stratification between the boreholes and buried objects along the proposed bridge alignment (Birid, 2016). Due to its importance and the low cost, such a survey is recommended for the marine infrastructure project. However, in order to acquire accurate survey output, a minimum of 2 m of water depth is recommended. The near-shore project with complex seabed conditions and shallow waters shall be treated with caution. A well-planned investigation methodology and equipment are crucial to successfully executing an investigation program as desired.

2.12 CONCLUSION

Non-invasive archaeological investigations require magnetometers and gradiometers to acquire high-resolution data. It may not be immediately apparent that there are distinct components of environmental and engineering subdisciplines within each 'problem' or application area. Applications such as groundwater and, especially, contaminant plume mapping, UXO, and archaeology are closely aligned with environmental geophysics. Here, the focus is on human health, society, and history, in contrast to the engineering properties and structures on which engineering geophysics focuses.

Geophysical methods also find applications primarily in the environmental subdiscipline, including pre-development baseline studies and re-development studies (i.e. brownfields). Applications to infrastructural development are of an engineering component, and in such problems, the detection and characterization of dangerous roadbed conditions underlying highways are made. This type of

application may involve detecting voids under roadways due to underground (i.e. mining or tunneling) excavations. Characterization of the relative integrity of reinforcing structures in bridges or other transportation structures is also carried out.

A very crucial application of the geophysical methods in the engineering subdiscipline of environmental and engineering geophysics is for dam safety; the safety of crucial infrastructures, like bridges, tunnels, nuclear and thermal plants, and multipurpose hydroelectric projects; and the safety of schools, colleges, vital administrative buildings, hospitals, and prisons, on which we determine the engineering properties of rocks and soils before construction is planned.

Geohazards studies cover all domains of constructing risk-resilient structures by applying all available tools of engineering geophysics that make society risk-resilient from both geogenic (landslides, earthquake, avalanches, tsunamis) and non-geogenic (hydrometeorological-induced) causes, such as glacial lake outburst flooding, or GLOF. Currently, it is very difficult to draw a clear distinction between each subdiscipline of environmental and engineering geophysics, since there are clear components overlapping with both environmental and engineering geophysics.

REFERENCES

- Ahmed, M.T., Woobaidullah, A.S.M., Quamruzzaman, C., Hossain, R., Islam, S. and Mehnaz, N., 2019. The evaluation of hydrogeological condition: A case study on Jessore Sadar, Bangladesh. *International Journal of Emerging Technology and Advanced Engineering* (ISSN 2250-2459, ISO 9001:2008 Certified Journal), vol. 9, no. 6, p. 26. Available at www.ijetae.com.
- Anderson, N.L., 2006. Selection of appropriate geophysical techniques: A generalized protocol based on engineering objectives and site characteristics. Proc., 2006 Highway Geophysics NDE Conference, pp. 29–47. Available at <http://2006geophysics.mst.edu/>.
- Birid, K.C. 2006. Geotechnical Investigations to assess Subsurface Stratigraphy of West Coast of Mumbai-A Case Study. Proceedings of 10th IAEG Congress, Geological Society of UK, Nottingham, United Kingdom, p. 6.
- Birid, K.C., 2016. Marine geotechnical and geophysical survey in complex geological formations. 4th International Conference on New Developments in Soil Mechanics and Geotechnical Engineering 2–4 June 2016, Near East University, Nicosia, North Cyprus, pp. 747–753.
- Bjerg, P.L., Tuxen, N., Reitzel, L.A., Albrechtsen, H.-J. and Kjeldsen, P., 2011. Natural attenuation processes in landfill leachate plumes at three Danish sites. *Ground Water*, vol. 49, no. 5, pp. 688–705.
- Carmichael, R.S. and Henry Jr., G., 1977. Gravity exploration for groundwater and bedrock topography. *Geophysics*, vol. 42, pp. 850–859.
- Chambers, J.E., Loke, M.H., Ogilvy, R.D. and Meldrum, P.I., 2004. Non-invasive monitoring of DNAPL migration through a saturated porous medium using electrical impedance tomography. *Journal of Contaminant Hydrology*, vol. 68, nos. 1–2, pp. 1–22.
- Christensen, L., Binning, P., Rosbjerg, D., Andersen, O. and Bauer-Gottwein, P., 2011. Using timelapse gravity for groundwater model calibration: An application to alluvial aquifer storage. *Water Resources Research*, vol. 47.
- Choudhury, K., Saha, D.K., Rai, M.K. and Naskar, D.C., 1997. Geophysical surveys for hydrogeological and environmental studies in the eastern Calcutta metropolis. *Indian Minerals*, vol. 51, nos. 1 & 2, pp. 41–56.
- Dahlin, T., Leroux, V. and Nissen, J., 2002. Measuring techniques in induced polarisation imaging. *Journal of Applied Geophysics*, vol. 50, pp. 279–298.
- Dev, H., Singh, R., Ramana, G.V. and Yadav, R.P., 2012. Determination of bearing capacity of gravel boulder deposit from footing load test. Proceedings of Indian Geotechnical Conference, Delhi (Paper no. A102).
- French, H. and Binley, A., 2004. Snowmelt infiltration: Monitoring temporal and spatial variability using time-lapse electrical resistivity. *Journal of Hydrology*, vol. 297, nos. 1–4, pp. 174–186.
- Gazoty, A., Behroozmand, A., Fiandaca, G. and Auken, E., 2011. Landfill characterization with geophysical methods: The Risby case study. *Applied Geophysics*.
- Herckenrath, D., 2012. Informing groundwater models with near-surface geophysical data. Ph.D. Thesis. DTU Environment Department of Environmental Engineering Technical University of Denmark.
- Huisman, J., Hubbard, S., Redman, J. and Annan, A., 2003. Measuring soil water content with ground penetrating radar: A review. *Vadose Zone Journal*, vol. 2, no. 4, pp. 476–491.

- Johnson, T.C., Rucker, D.F. and Glaser, D.R., 2015a. Near-surface geophysics at the Hanford Nuclear site, the United States. In: *Treatise on Geophysics*. 2nd edn. vol. 11. St. Louis, MO: Elsevier, pp. 571–595, DOI: 10.1016/b978-0-444-53802-4.00205-0.
- Johnson, T.C., Versteeg, R.J., Day-Lewis, F.D., Major, W. and Lane, J.W., 2015b. Time-lapse electrical geophysical monitoring of amendment-based biostimulation. *Groundwater*, vol. 53, no. 6, pp. 920–932, DOI: 10.1111/gwat.12291.
- Kjeldsen, P., Barlaz, M.A., Rooker, A.P., Baun, A., Ledin, A. and Christensen, T.H., 2002. Present and long-term composition of MSW landfill leachate: A review. *Critical Reviews in Environmental Science and Technology*, vol. 32, no. 4, pp. 297–336.
- Knight, R., 2001. Ground penetrating radar for environmental applications. *Annual Review of Earth and Planetary Sciences*, vol. 29, pp. 229–255.
- Kowalsky, M.B., Finsterle, S., Peterson, J., Hubbard, S., Rubin, Y., Majer, E., Ward, A. and Gee, G., 2005. Estimation of field-scale soil hydraulic and dielectric parameters through joint inversion of GPR and hydrological data. *Water Resources Research*, vol. 41, no. 11.
- Lambot, S., Slob, E., Rhebergen, J., Lopera, O., Jadoon, K.Z. and Vereecken, H., 2009. Remote estimation of the hydraulic properties of a sand using full-waveform integrated hydrogeophysical inversion of time-lapse, off-ground GPR data. *Vadose Zone Journal*, vol. 8, no. 3, pp. 743–754.
- Legchenko, A., Baltassat, J.M., Beauce, A. and Bernard, J., 2002. Nuclear magnetic resonance as a geophysical tool for hydrogeologists. *Journal of Applied Geophysics*, vol. 50, nos. 1–2, pp. 21–46.
- Legchenko, A. and Valla, P., 2002. A review of the basic principles for proton magnetic resonance sounding measurements. *Journal of Applied Geophysics*, vol. 50, nos. 1–2, pp. 3–19.
- Leroy, P. and Revil, A., 2009. A mechanistic model for the spectral induced polarization of clay materials. *Journal of Geophysical Research-Solid Earth*, vol. 114.
- Lesmes, D.P. and Friedman, S.P., 2005. Relationships between the Electrical and Hydrogeological Properties of Rocks and Soils. *Water Science and Technology Library*, vol. 50, Part 2, pp. 87–128.
- Looms, M.C., Jensen, K.H., Binley, A. and Nielsen, L., 2008. Monitoring unsaturated flow and transport using cross-borehole geophysical methods. *Vadose Zone Journal*, vol. 7, no. 1, pp. 227–237.
- Lorah, M.M., Cozzarelli, I.M. and Boehlke, J., 2009. Biogeochemistry at a wetland sediment-alluvial aquifer interface in a landfill leachate plume. *Journal of Contaminant Hydrology*, vol. 105, nos. 3–4, pp. 99–117.
- Meju, M.A., 2000. Geoelectrical investigation of old/abandoned, covered landfill sites in urban areas: Model development with a genetic diagnosis approach. *Journal of Applied Geophysics*, vol. 44, nos. 2–3, pp. 115–150.
- Michot, D., Dorigny, A. and Benderitter, Y., 2003. Determination of water flow direction and corn roots-induced drying in an irrigated Beauce CALCISOL, using electrical resistivity measurements. *Comptes Rendus de l'Academie des Sciences Serie II Fascicule A Sciences de la Terre et des Planetes*, vol. 332, no. 1, pp. 29–36.
- Miller, R.D., Steeples, D.W. and Brannan, M., 1989. Mapping a bedrock surface under dry alluvium with shallow seismic reflections. *Geophysics*, vol. 54, no. 12, pp. 1528–1534.
- Milosevic, N., Thomsen, N.I., Juhler, R.K., Albrechsten, H.-J. and Bjerg, P.L., 2012. Identification of discharge zones and quantification of contaminant mass discharges into a local stream from a landfill in a heterogeneous geologic setting. *Journal of Hydrology*, vol. 446–447, pp. 13–23.
- Mohnke, O. and Yaramanci, U., 2008. Pore size distributions and hydraulic conductivities of rocks derived from magnetic resonance sounding relaxation data using multi-exponential decay time inversion. *Journal of Applied Geophysics*, vol. 66, nos. 3–4, pp. 73–81.
- Montgomery, E.L., 1971. Determination of specific yield using gravity measurements. *Transactions-American Geophysical Union*, vol. 52, no. 4, p. 205.
- Naskar, D.C. and Das, L.K., 2009. Resistivity survey across the Narmada-Son lineament and along the Sanwer-Borgaon traverse, Madhya Pradesh, Central India. *Bollettino di Geofisica Teorica ed Applicata*, vol. 50, no. 4, pp. 305–318.
- Naudet, V., Revil, A., Rizzo, E., Bottero, J.Y. and Begassat, P., 2004. Groundwater redox conditions and conductivity in a contaminant plume from geoelectrical investigations. *Hydrology and Earth System Sciences*, vol. 8, no. 1, pp. 8–22.
- Pelton, W.H., Ward, S.H., Hallof, P.G., Sill, W.R. and Nelson, P.H., 1978. Mineral discrimination and removal of inductive coupling with multifrequency IP. *Geophysics*, vol. 43, pp. 588–609.
- Revil, A. and Florsch, N., 2010. Determination of permeability from spectral induced polarization in granular media. *Geophysical Journal International*, vol. 181, no. 3, pp. 1480–1498.

- Seguin, M.K., Gupta, O.P., Nagrajan, V., Jain, A.K., Sen, N., Mahapatra, S. and Mishra, O.P., 1994. Identification and delineation of abandoned, unknown and surveyed galleries in Raniganj coalfield by geophysical methods, Exploration Technology for modern coal mines *Coal Exploration. Seminar*, vol. VI-49–VI-70.
- Slater, L.D., 2007. Near surface electrical characterization of hydraulic conductivity: From petrophysical properties to aquifer geometries—a review. *Surveys in Geophysics*, vol. 28, nos. 2–3, pp. 169–197.
- Slater, L.D., 2022. Geophysical methods. This page was last edited on 26 April 2022, at 22:00. Available at https://www.envirowiki.org/index.php?title=Geophysical_Methods_Case_Studies&oldid=15826.
- Slater, L.D., Ntarlagiannis, D., Day-Lewis, F.D., Mwakanyamale, K., Versteeg, R.J., Ward, A., Strickland, C., Johnson, C.D. and Lane, J.W., 2010. Use of electrical imaging and distributed temperature sensing methods to characterize surface water—groundwater exchange regulating uranium transport at the Hanford 300 Area, Washington. *Water Resources Research*, vol. 46, p. W10533.
- Sogade, J.A., Scira-Scappuzzo, F., Vichabian, Y., Shi, W., Rodi, W., Lesmes, D.P. and Morgan, F.D., 2006. Induced-polarization detection and mapping of contaminant plumes. *Geophysics*, vol. 71, no. 3, pp. B75–B84.
- Telford, W.M., Geldari, L.P. and Sheriff, R.E., 1990. *Applied Geophysics: Prospecting Geophysical Methods*. Cambridge and New York: Cambridge University Press, p. 770.
- Thomsen, R., Sondergaard, V.H. and Sorensen, K.I., 2004. Hydrogeological mapping as a basis for establishing site-specific groundwater protection zones in Denmark. *Hydrogeology Journal*, vol. 12, no. 5, pp. 550–562.
- Topp, G.C., Davis, J.L. and Annan, A.P., 1980. Electromagnetic determination of soil-water content—measurements in coaxial transmission-lines. *Water Resources Research*, vol. 16, no. 3, pp. 574–582.
- van Gelderen, M., Haagmans, R. and Bilker, M., 1999. Gravity changes and natural gas extraction in Groningen. *Geophysical Prospecting*, vol. 47, pp. 979–993.
- Vaudelet, P., Revil, A., Schmutz, M., Franceschi, M. and Begassat, P., 2011. Induced polarization signatures of cations exhibiting differential absorption behaviors in saturated sands. *Water Resources Research*, vol. 47.
- Vouillamoz, J., Favreau, G., Massuel, S., Boucher, M., Nazoumou, Y. and Legchenko, A., 2008. Contribution of magnetic resonance sounding to aquifer characterization and recharge estimate in semiarid Niger. *Journal of Applied Geophysics*, vol. 64, nos. 3–4, pp. 99–108.
- Williams, K.H., Kemna, A., Wilkins, M.J., Druhan, J., Arntzen, E., N'Guessan, A.L., Long, P.E., Hubbard, S.S. and Banfield, J.F., 2009. Geophysical monitoring of coupled microbial and geochemical processes during stimulated subsurface bioremediation. *Environmental Science & Technology*, vol. 43, no. 17, pp. 6717–6723.
- Zawila, J.S., Damiata, B.N., Biehler, S.C. and Lee, T.C., 1997. Gravity mapping of subsurface structures in batholithic terrain. Symposium on the Application of Geophysics to Engineering and Environmental Problems, pp. 801–810, Reno.

3 Potential Field Methods (Gravity and Magnetic)

Gravity and magnetics geophysical methods, also known as the ‘potential fields’ methods in geophysical exploration, are used in natural resource exploration to help locate mineral, petroleum, and groundwater resources. They are an indirect (aboveground) method of sensing the density and magnetization associated with the magnetic susceptibility of the subsurface geological formations as the physical properties contrast between the host and the target surface materials for the purpose of locating faults and other lithology. These geophysical methods are cheaper and can investigate larger areas of ground quickly and authentically. These potential geophysical methods are often used to identify smaller target areas in the field. Most of the time, density contrasts prevail at boundaries between diverse geological formations, but man-made boundaries, such as tunnels and mines, also represent contrasts. It has been documented by several researchers that gravity variations caused by density inhomogeneities are relatively too small ($\geq 10^{-6}$ of the Earth’s gravity), which are accurately measured using extremely sensitive gravimeters. These two potential methods have wide-scale applicability in conducting both reconnaissance and detailed surveys of the areas or regions when subjected to geophysical exploration for specific purposes. The gravity surveying closely with the magnetic method is because of the relative prominence of the respective physical property contrasts (density and magnetic susceptibility) associated with various geological formations, besides both requiring conceptually similar interpretation techniques.

Gravity and magnetic potentials bear a good relationship based on Poisson’s theorem, which states that magnetic potential (B) is proportional to the partial derivative of the gravitational potential (U) with respect to the orientation direction (S) of the same source density, ‘ ρ ’, and the intensity of magnetization (I) with a proportionality constant, ‘ C ’, with its numerical value of 10^{-7} in SI units, which can be expressed as follows:

$$B = C \cdot \frac{I}{G\rho} \cdot \frac{\delta U}{\delta S} \quad (1)$$

Where G is the universal gravitational constant.

If the direction of magnetization (I) is vertical ($s = z$), as it is in high-latitude areas, the magnetic potential (B) can be calculated directly from the acceleration due to gravity (g), which can be expressed as $\frac{\delta U}{\delta z}$. Under such circumstances, the magnetic field component in the vertical direction ($B_z = \frac{\delta B}{\delta z}$) can be estimated from the vertical derivative of gravity by Poisson’s relationship, as follows:

$$B_z = C \cdot \frac{I}{G\rho} \cdot \frac{\delta g}{\delta z} \quad (2)$$

These equations as mentioned under (1) and (2) provide a clue for making estimates of gravity and magnetic fields from each other by assuming a common source, which also suggests that the magnetic field varies one power faster with respect to distance than does gravity, which in turn suggests that the magnetic field is proportional to the rate of spatial change of gravity. Accordingly, the magnetic field, due to the source, will be more sensitive to depth and will better depict the location of the source than will gravity field from the same source (Sharma, 1997).

3.1 GRAVITY METHOD

The gravity field on the surface of the Earth is not uniformly the same everywhere. It varies with the distribution of the mass materials below. This lateral change can be measured and interpreted in terms of likely causative geology. A *gravity survey* is regarded as an indirect (surface) method of calculating the density property of subsurface materials. The higher the gravity values correspond, the denser the rock beneath. In gravity prospecting, subsurface geology is investigated based on the Earth's gravitational field generated by differences in density between subsurface rocks. A very wide range of geological situations gives rise to such differences in density producing significant gravity anomalies, for example, buried valley, granite plutons, sedimentary basins, salt domes, massive sulphides, etc. Gravity methods have been used most extensively in the search for oil and gas, particularly in the 20th century. While such methods are still employed very widely in hydrocarbon exploration, many other applications have been found, some examples of which are (Reynolds, 1997a, 1997b) listed in what follows (Table 3.1).

Most recently, the National Geophysical Mapping Programme (NGPM) of the GSI aims to generate baseline geophysical data to help locate and delineate subsurface geological structures and develop conceptual models of mineralized belts. This is done by a combination of geological and geochemical maps to locate potential/probable target areas for mineral exploration. Thus, the gravity–magnetic anomaly maps help in planning and formulating strategies for structural and tectonic studies, geotechnical studies, mitigation of natural hazards, and management of the environment and urban planning. So far, NGPM has covered a total of 0.28 million square kilometre area, and this is about 9% of the land area (GSI, 2016).

3.2 KEPLER'S LAWS OF PLANETARY MOTION

Kepler took many years to fit the observations of Tycho Brahe into three laws of planetary motion which used to derive Newton's law of gravitation. The first and second laws were published in 1609, and the third law appeared in 1619. The laws may be formulated as follows.

- First Law (Kepler's Law of Elliptical Orbits, 1609)

Every planet moves around the sun in a closed elliptical orbit, with the sun at one of its foci.

TABLE 3.1

The Range of Applications for Gravity Surveying (Independent of Instrument Type)

Methodology	Applications
Gravity method	Hydrocarbon exploration
	Regional geological studies
	Isostatic compensation determination
	Exploration for, and mass estimation of, mineral deposits
	Detection of subsurface cavities (microgravity)
	Location of buried rock valleys
	Determination of glacier thickness
	Tidal oscillations
	Archaeogeophysics (microgravity), for example, location of tombs
	Shape of the Earth (geodesy)
	Military (especially for missile trajectories)
	Monitoring volcanoes

The following figure (Figure 3.1a; Wikipedia) shows the path of the planet around the Sun. Kepler's law of elliptical orbits explains that due to elliptical motion and the sun at one of the foci, the distance of a planet from the sun changes continuously. The closest point on the orbit of the planet from the sun is 'B'. This closest point on the orbit of the planet to the sun is called *perihelion*, and this minimum distance is called *perigee*. Thus, $\text{perigee} = SB = a - ae = a(1 - e)$. Where a is the semi-major axis and e is the eccentricity of the ellipse. The farthest point on the orbit of the planet from the sun is A. This farthest point on the orbit of the planet to the sun is called *aphelion*, and this maximum distance is called *apogee*. Thus, $\text{apogee} = SA = a + ae = a(1 + e)$.

- Second Law (Kepler's Law of Equal Areas, 1609)

The radius vector drawn from the sun to the planet sweeps out equal areas in equal time, that is, aerial velocity of the radius vector is constant.

Thus, the aerial velocity of the satellite is always constant.

$$\text{Namely, } \frac{A}{t} = \text{constant} \quad (3.1)$$

In this case, the planet covers unequal distances in equal time. Thus, the planet has variable speed. When the planet is very close to the sun, the maximum distance is covered in a given time (Figure 3.1b; Wikipedia). This explains the fact that the planet has maximum velocity when it is at perihelion. Hence, it has maximum kinetic energy at perihelion. When the planet is very far from the sun, the minimum distance is covered in a given time. Thus, the planet has minimum velocity when it is at aphelion. Hence, it has minimum kinetic energy at perihelion.

- Kepler's Third Law (Kepler's Law of Period, 1618)

The square of the period of a satellite is directly proportional to the cube of the semi-major axis of its elliptical orbit.

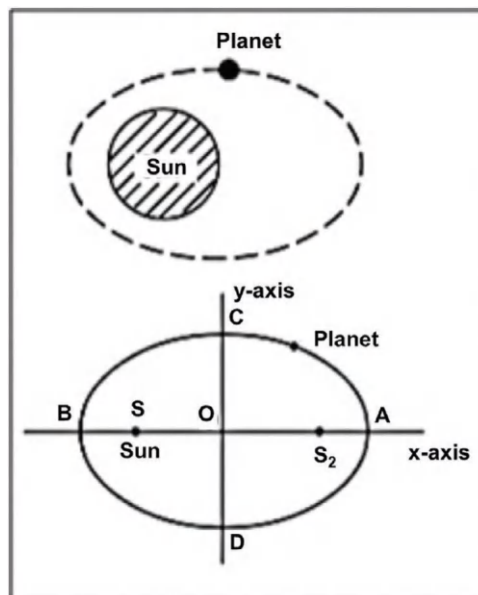


FIGURE 3.1A First law of Kepler.

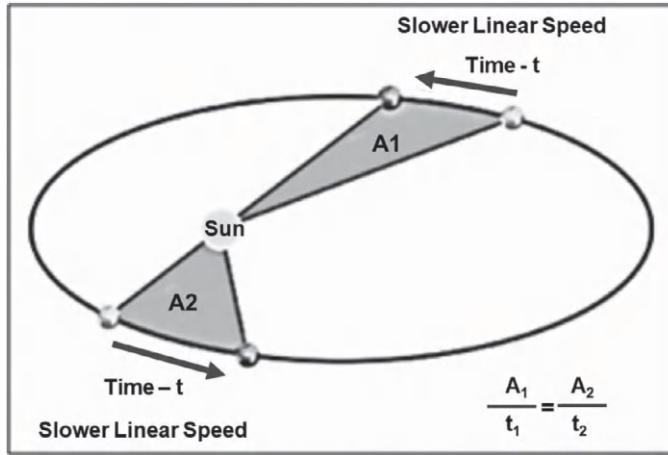


FIGURE 3.1B Kepler's second law.

If T is the period of the satellite and r is the semi-major axis or the radius of the circular orbit of the satellite,

$$T^2 \propto r^3$$

In general,

$$\frac{T_1^2}{T_2^2} = \frac{r_1^3}{r_2^3} \quad (3.2)$$

A similar result is obtained for the elliptical orbit and if the radius r of the circular orbit is replaced by semi-major axis a . Kepler's laws are empirical laws because they are based on observations and not on theory. Kepler's laws are applicable whenever an inverse square law is involved. These are applicable to the solar system as well as to the artificial satellite systems.

3.3 BASIC PRINCIPLES

The basis of the method is Newton's law of gravitation described in *Principia Mathematica* in 1687. One of the basic forces of nature is the attraction between all masses. This attraction is called the force of gravity. The force of attraction (gravitational force) between two point masses m_1 and m_2 at a distance r is given by

$$F = -G \frac{m_1 m_2}{r^2} \hat{r}_1 \quad (3.3)$$

A negative sign is put because the force is an attraction. F is a force in dynes, $\hat{r}_1$ is a unit vector directed from m_1 to m_2 (masses in gms), and r is the distance between m_1 and m_2 (in centimetres) (Figure 3.2; Saibi, 2018). Figure 3.3 shows the gravitational attraction of a small mass on Earth (Saibi, 2018).

Then, G = gravitational constant = 6.67×10^{-8} dyne-cm²/gm², which is the force in dynes between two small uniform spheres of 1 gm each separated by 1 cm. In SI (International System) unit, $G = 6.67 \times 10^{-11}$ N-m²/kg². The dimension of G is $L^3 M^{-1} T^{-2}$.

Consider the gravitational attraction of a spherical, non-rotating, homogeneous Earth of mass M and radius R on a small mass m on its surface. As the mass of a sphere acts as though it were concentrated at the centre of the sphere,

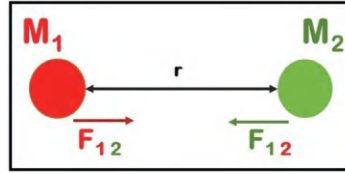


FIGURE 3.2 Attractive force acting between two bodies.

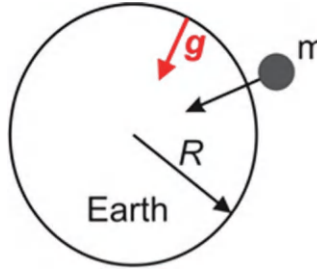


FIGURE 3.3 The gravitational attraction of a small mass m on Earth.

$$mg = \frac{GMm}{R^2} \quad (3.4)$$

$$g = \frac{GM}{R^2} \quad (3.5)$$

where g is the acceleration due to gravity (*acceleration due to gravity* is the acceleration gained by an object due to the gravitational force; its SI unit is m/s^2 and is a vector quantity since it has both magnitude and direction, and the standard value of g on the surface of the Earth at sea level is 9.8 m/s^2), or simply gravity (*gravity* is the force with which Earth attracts a body towards its centre), which is the acceleration imparted to a freely falling body in the absence of frictional forces. The gravity method takes care of measuring the gravitational attraction exerted by the Earth at a measurement station on the surface. It is observed that the strength of the gravitational field is directly proportional to the mass, and thereby the density, of subsurface materials. Anomalies in the Earth's gravitational field result from lateral variations in the density of subsurface materials and the distance to these bodies from the measuring equipment (Mariita, 2009).

3.3.1 UNITS OF GRAVITY

The mean value of gravity on the Earth's surface is about 980 cm/sec^2 , or 980 Gals ($1 \text{ Gal} = 1 \text{ cm/sec}^2$). In SI unit, the gravity at the Earth's surface is 9.8 m/sec^2 . For geologic and geodetic studies, the milligal (mGal , $\frac{1}{1000} \text{ Gal}$) has a common unit for expressing gravitational acceleration. Another unit, less commonly used, is the gravity unit:

$$\text{gu} = 10^{-6} \text{ m/sec}^2 = 10^{-4} \text{ cm/sec}^2 = 10^{-4} \text{ Gal} = 10^{-1} \text{ mGal} = 0.1 \text{ mGal} = 100 \text{ } \mu\text{Gal}$$

Variations of gravity caused by density variations in the subsurface are of the order of 100 mGal. The accuracy of gravity survey on land is generally $\pm 0.1 \text{ mGal}$.

3.3.2 MASS OF THE EARTH

$$M = \frac{gR^2}{G} \quad (3.6)$$

Taking $R = 6,400$ km, $G = 6.67 \times 10^{-8}$ c.g.s. unit, and $g = 981$ cm/sec², we get

$$M = 6.1 \times 10^{27} \text{ gms}$$

On such an Earth, gravity would be constant. However, the Earth's ellipsoidal shape, rotation, irregular surface relief, and internal mass distribution cause gravity to vary from place to place. This law is applicable only to particles or to bodies whose dimensions are small compared to the distance of separation.

3.3.3 GRAVITATIONAL POTENTIAL

Gravitational potential at a point in a given field is defined as the work done by the force in moving a unit mass from an arbitrary reference point (usually at an infinite distance) to the point in question. The gravitational field is most usefully defined in terms of gravitational potential U .

$$U = \int_{\infty}^r Gm_1 \frac{dr}{d^2} = -\frac{Gm_1}{r} \quad (3.7)$$

Two masses—one of unit magnitude, and the other of magnitude m_1 —are initially an infinite distance apart. The first derivative of U in any direction gives the value of gravity in that direction.

$$g_z = \frac{\partial U}{\partial z} \quad (3.8)$$

Gravitation attraction, or gravity, is a vector quantity with both magnitude and direction (vertically downward). The gravitation potential U is a scalar quantity having magnitude only. An *equipotential surface* is defined as one on which U is constant. The sea-level surface, or geoid, is the most easily recognized geopotential surface, which is horizontal everywhere and perpendicular to the direction of gravity.

3.3.3.1 Acceleration and Potential of a Distribution of Mass

Until now, we have considered only the gravitational acceleration and potential of point masses. A solid body may be considered to be composed of numerous small particles, each of which exerts a gravitational attraction at an external point P (Figure 3.4a). To calculate the gravitational acceleration of the object at the point P , we must form a vector sum of the contributions of the individual discrete particles (Lowrie, 2007). Each contribution has a different direction. Assuming m_i to be the mass of the particle at distance r_i from P , this gives an expression like

$$g = -G \frac{m_1}{r_1^2} \hat{r}_1 - G \frac{m_2}{r_2^2} \hat{r}_2 - G \frac{m_3}{r_3^2} \hat{r}_3 - \dots$$

Depending on the shape of the solid, this vector sum can be quite complicated.

An alternative solution to the problem is found by first calculating the gravitational potential and then differentiating it, as in eq. 3.3, to get the acceleration. The expression for the potential at P is

$$U = -G \frac{m_1}{r_1} - G \frac{m_2}{r_2} - G \frac{m_3}{r_3} - \dots$$

This is a scalar sum, which is usually simpler to calculate than a vector sum.

3.3.3.1.1 Newtonian or Three-Dimensional Potential

Newtonian or three-dimensional potential put forth the condition that if the density of the matter in each volume is known, the mass of the small element can be calculated and its contribution to the potential at the external point P can be determined. By integrating over the volume of the body, its gravitational potential at P can be calculated. At a point in the body with coordinates $r(x, y, z)$, let the density be $\rho(x, y, z)$, and let its distance from P be $r(x, y, z)$, as in Figure 3.4b. The potential due to an element of mass dm at a distance r from a point P is (Lowrie, 2007)

$$dU = \frac{Gdm}{r} = \frac{G\sigma dx dy dz}{r}$$

where $r^2 = x^2 + y^2 + z^2$, and σ is the density.

The potential of the total mass m will be

$$U = G\sigma \iiint \frac{1}{r} dx dy dz$$

The acceleration in the direction of z -axis is

$$g_z = \frac{\partial U}{\partial z} = G\sigma \iiint \frac{z}{r^3} dx dy dz \quad (3.9)$$

3.3.3.1.2 Logarithmic or Two-Dimensional Potential

For the 2-D body, which is very long in the y -direction, the gravity attraction has to be derived from logarithmic potential rather than Newtonian.

$$U = G\sigma \iint dx dz \int \frac{dy}{r} = 2 G\sigma \iint \log \left(\frac{1}{r} \right) dx dz$$

The gravity effect for the 2-D body is

$$g_z = \frac{\partial U}{\partial z} = 2G\sigma \iint \frac{z}{r^2} dx dz \quad (3.10)$$

3.3.3.1.3 Potential Field Equations

Gauss's divergence theorem states that the integral of the divergence of a vector field over a region of space is equivalent to the integral of the outward normal component of the field over the surface enclosing the region.

$$\int \partial \cdot \mathbf{g} dV = \int \mathbf{g}_n \cdot d\mathbf{S}$$

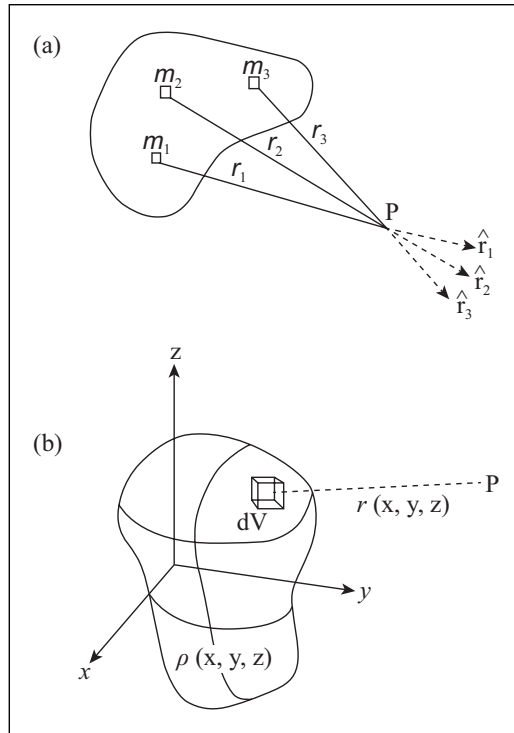


FIGURE 3.4 (a) Each small particle of a solid body exerts a gravitational attraction in a different direction at an external point P . (b) Computation of the gravitational potential of a continuous mass distribution.

If there is no attracting material within this volume, the integral is 0, and

$$\begin{aligned} \partial \cdot \mathbf{g} &= 0 \\ \text{Or, } \partial \cdot \partial U &= \partial^2 U = 0 \end{aligned}$$

$$\text{Where } \partial^2 U = \partial^2 U / \partial x^2 + \partial^2 U / \partial y^2 + \partial^2 U / \partial z^2.$$

But if there is a particle of mass m within the volume, say, at the centre of the spherical surface, with radius r , then

$$\int \mathbf{g}_n \cdot d\mathbf{S} = \frac{Gm}{r^2} (4\pi r^2) = 4\pi Gm$$

If the total mass inside is M , then

$$\begin{aligned} \int \partial \cdot \mathbf{g} \, dV &= \int \mathbf{g}_n \cdot d\mathbf{S} = 4\pi GM \\ \partial \cdot \mathbf{g} &= 4\pi G\sigma, \text{ for very small volume covering point mass} \\ \text{Or, } \partial^2 U &= 4\pi G\sigma \end{aligned} \tag{3.11}$$

Which is Poisson's equation. This shows that gravity potential satisfies Laplace's equation in free space and Poisson's equation in an attracting medium.

3.3.4 GRADIENTS OF GRAVITY

The rate of change of gravity with distance is designated as the *gradient*. Vertical gradient of gravity for 3-D case is

$$\frac{\partial g_z}{\partial z} = \frac{\partial^2 U}{\partial z^2} = G\sigma \iiint \left(\frac{1}{r^3} - \frac{3z}{r^5} \right) dx dy dz$$

For the 2-D case, it is

$$\frac{\partial g_z}{\partial z} = \frac{\partial^2 U}{\partial z^2} = 2G\sigma \iint \left(2z^2 - \frac{r^2}{r^4} \right) dx dz$$

3.3.5 SECOND DERIVATIVE

The second derivative for the 3-D case is

$$\frac{\partial^2 g_z}{\partial z^2} = \frac{\partial^3 U}{\partial z^3} = 3G\sigma \iiint \left(\frac{5z^3}{r^7} - \frac{3z}{r^5} \right) dx dy dz$$

For the 2-D case, it is

$$\frac{\partial^2 g_z}{\partial z^2} = \frac{\partial^3 U}{\partial z^3} = 4G\sigma \iint \left(\frac{3z}{r^4} - \frac{4z^3}{r^6} \right) dx dz$$

3.3.6 VERTICAL GRADIENT OF GRAVITY FROM TOP AND BOTTOM OF TOWER

$$\text{observed gradient} = \frac{g_1 - g_2}{d}$$

Theoretical gradient = 0.3086 mGal/m, which is the vertical gradient of gravity at sea level. The maximum deviation from this value at any latitude is −0.0002 mGal/m.

3.3.7 HORIZONTAL GRADIENT OF GRAVITY FROM TOP AND BOTTOM OF TOWER

This is obtained by taking derivatives of g_z along the x and y directions. For the x component, for the 3-D case:

$$U_{xz} = \frac{\partial^2 U}{\partial x \partial z} = 3G\sigma \iiint \frac{xz}{r^5} dx dy dz$$

For the 2-D case:

$$U_{xz} = \frac{\partial^2 U}{\partial x \partial z} = 4G\sigma \iint \frac{xz}{r^4} dx dz$$

This is a very significant parameter in gravity because the sharpness of a profile is an indication of the depth of the anomalous mass.

3.3.8 SHAPE OF THE EARTH

It is necessary to know the shape of the Earth to calculate the gravity field at any point on it. With an equatorial radius of 6,378 km and a polar radius of 6,357 km, the Earth is considered to be a spheroid of reference rather than a simple sphere (Figure 3.5 to Figure 3.7) (Strahler, 1963; Reynolds, 1998). The geometrical flattening f , or $\frac{R_e - R_p}{R_e}$, and the gravitational flattening, $\frac{g_p - g_e}{g_e}$, are related to each other by Clairaut's theorem, which states

$$\frac{R_e - R_p}{R_e} + \frac{g_p - g_e}{g_e} = \frac{5}{2} \frac{R_e \omega^2}{g_e}$$

where R and g are the radii and gravity, with suffixes e and p standing for 'equator' and 'pole', and $R_e \omega^2$ is the centrifugal acceleration (Reynolds, 1998).

Modern analyses of the Earth's shape depend on the precise observations of the orbits of artificial Earth satellites. These data are used to define a best-fitting oblate ellipsoid, called the international reference ellipsoid. In 1930, geodesists and geophysicists defined an optimum reference ellipsoid based on the best available data at that time. The dimensions of this figure have been subsequently refined as more exact data have become available. In 1980, the International Association of Geodesy adopted a geodetic reference system (GRS80) in which the reference ellipsoid has an equatorial radius R_e equal to 6,378.137 km and a polar radius R_p equal to 6,356.752 km. Subsequent determinations have resulted in only minor differences in the most important geodetic parameters. The radius of the equivalent sphere R is found from $R = (R_e^2 R_p)^{\frac{1}{3}}$ to be 6,371.000 km. Compared to the best-fitting sphere, the spheroid is flattened by about 14.2 km at each pole, and the equator bulges by about 7.1 km.

The recent estimate of the flattening f of the optimum reference ellipsoid defined in 1930 was exactly $\frac{1}{297}$. This ellipsoid, and the variation of gravity on its surface, served as the basis of gravimetric surveying for many years, until the era of satellite geodesy and highly sensitive gravimeters showed it to be too inexact. A recent best estimate of the flattening is $f = 3.35287 \times 10^{-3}$ (i.e. $f = (\frac{1}{298.252})$).

Centrifugal acceleration is the principal cause of the polar flattening due to the deformation effect. This is maximum at the equator, where gravitational acceleration is smallest. As a result of the flattening, the distribution of mass within the Earth is not simply dependent on radius. The moments of inertia of the Earth about the rotation axis (C) and any axis in the equatorial plane (A) are unequal. The principal moments of inertia define the *dynamical ellipticity*:

$$H \approx \frac{C - A}{C} = \frac{1}{305.457} = \text{current optimum value}$$

The ellipsoidal shape of the Earth resulting from its rotation has important consequences, not only for the variation with latitude of gravity on the Earth's surface, but also for the Earth's rate of rotation and the orientation of its rotational axis. These are modified by torques that arise from the gravitational attractions of the sun, the moon, and the planets on the ellipsoidal shape. If the Earth is assumed to be a rotating fluid in perfect hydrostatic equilibrium (as assumed by Newton's theory),

the flattening should be $\frac{1}{299.5}$, slightly smaller than the observed value.

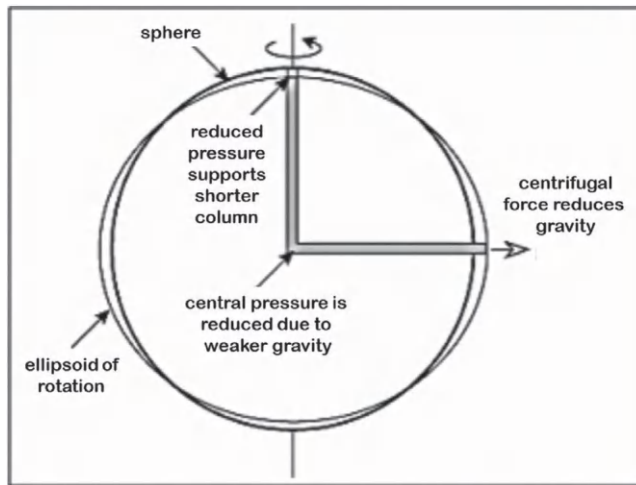


FIGURE 3.5 Newton's argument that the shape of the rotating Earth should be flattened at the poles and should bulge at the equator was based on hydrostatic equilibrium between polar and equatorial pressure columns.

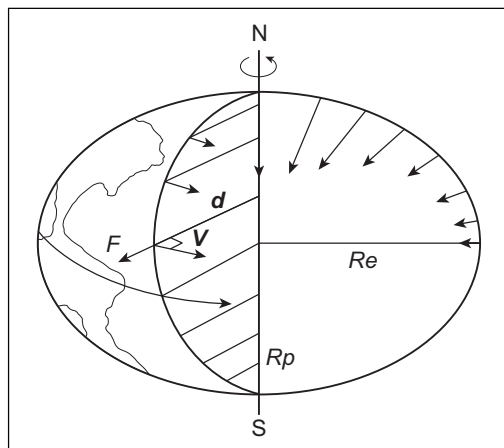


FIGURE 3.6 Centrifugal acceleration and variation of gravity with latitude ϕ (not to scale). Gravity is 5,186 mGal greater at the poles than at the equator. The acceleration due to gravity varies with latitude due to two effects: the Earth's shape and the Earth's rotation.

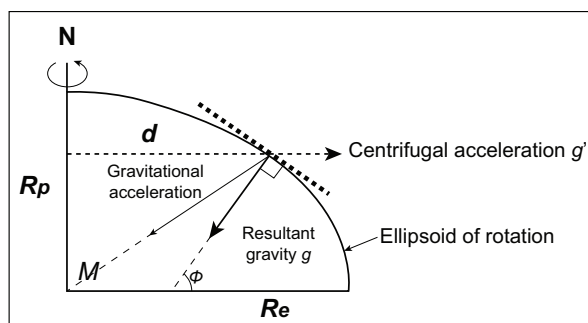


FIGURE 3.7 Resultant of centrifugal acceleration g' and the acceleration due to gravity g .

3.3.9 SPHEROID AND GEOID

The *spheroid* of reference is a mathematical surface coinciding with the mean sea level, and the value of g at any point on it is defined by the following formula adopted by the International Association of Geodesy in 1967, replacing the international formula of 1930. Clairaut's gravity formula relates gravity to latitude on the reference spheroid according to an equation of the form

$$g_{\phi} = g_0 (1 + k_1 \sin^2 \phi + k_2 \sin^2 2\phi)$$

where g_{ϕ} is the predicted value of gravity at a latitude, g_0 is the value of gravity at the equator, and k_1 and k_2 are constants, dependent on the shape and speed of rotation of the Earth. According to the 1930 formula (Cassinis, 1930):

$$g_{\phi} = 978.049 (1 + 0.0052884 \sin^2 \phi - 0.0000059 \sin^2 2\phi) \text{ Gal}$$

And according to the 1967 formula (Heiskanen and Moritz, 1967):

$$g_{\phi} = 978.0318 (1 + 0.0053024 \sin^2 \phi - 0.0000059 \sin^2 2\phi) \text{ Gal}$$

$$g_{1967} - g_{1930} = (-17.2 + 13.6 \sin^2 \phi) \text{ mGal}$$

However, the gravity relation given earlier is only approximate because of warping in the equipotential surface of gravity. In general terminology, the actual equipotential surface of the Earth that corresponds to sea level is called the geoid. It reflects the true distribution of mass inside the Earth and differs from the theoretical ellipsoid by small amounts (Kearey et al., 2002). The geoid agrees with the free ocean surface located far from the land, excluding the temporary perturbing effects of tides and winds. Over the continents, the geoid is affected by the mass of land above mean sea level (Figure 3.8a). A series of experimental studies revealed that the mass within the ellipsoid causes a downward gravitational attraction toward the centre of the Earth, but a hill or mountain, whose centre of gravity is outside the ellipsoid, causes an upward attraction. This causes a local elevation of the geoid above the ellipsoid. The displacement between the geoid and the ellipsoid is called a geoid undulation; the elevation caused by the mass above the ellipsoid is a positive undulation.

In computing the theoretical figure of the Earth, the distribution of mass beneath the ellipsoid is assumed to be homogeneous. A local excess of mass under the ellipsoid will deflect and strengthen gravity locally. The potential of the ellipsoid is achieved further from the centre of the Earth. The equipotential surface is forced to warp upward while remaining normal to gravity (Kearey et al., 2002). This gives a positive geoid undulation over a mass excess under the ellipsoid (Figure 3.8b). Conversely, negative geoid undulations are caused by mass deficit beneath the ellipsoid, which will deflect the geoid below the ellipsoid. Consequently, with the uneven topography and heterogeneous internal mass distribution of the Earth, the geoid is a bumpy equipotential surface.

A combination of satellite data and surface gravity measurements was used to construct the Goddard Earth Model (GEM) 10. A global comparison between a reference ellipsoid with flattening $\frac{1}{298.257}$ and the geoid surface computed from the GEM 10 model shows long-wavelength geoid undulations (Lerch et al., 1979). The largest negative undulation (−105 m) is in the Indian Ocean south of India, and the largest positive undulation (+ 73 m) is in the equatorial Pacific Ocean north of Australia. These large-scale features are too broad to be ascribed to shallow crustal or lithospheric mass anomalies. They are thought to be due to heterogeneities that extend deep into the lower mantle, but their origin is not yet understood.

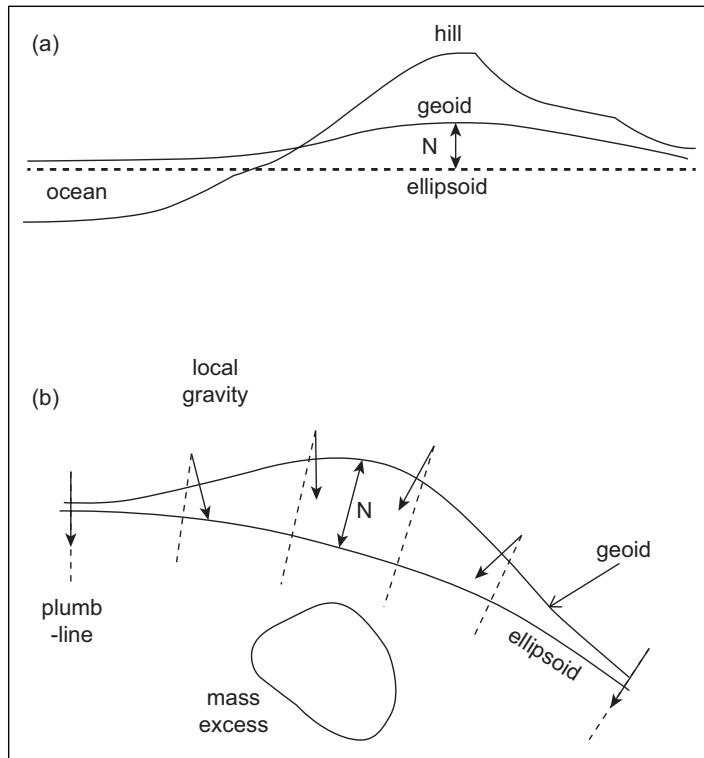


FIGURE 3.8 (a) A mass outside the ellipsoid or (b) a mass excess below the ellipsoid elevates the geoid above the ellipsoid. N is the geoid undulation.

3.3.10 VARIATION OF G FROM PLACE TO PLACE

3.3.10.1 Variation in Acceleration Due to Gravity Due to Altitude

Above the surface of the Earth (Figure 3.9a; Wikipedia), the gravitational force

$$F = \frac{GMm}{(R + h)^2} = mg$$

$$g = \frac{GM}{(R + h)^2}$$

That is, it is inversely proportional to $(R + h)^2$.

Therefore, g will be less and less as the station height h increases from the Earth's surface.

3.3.10.2 Variation in Acceleration Due to Gravity Due to Depth

Below the surface of the Earth (Figure 3.9b; Wikipedia), the gravitational force

$$F = \frac{GMm}{(R - d)^2} = mg$$

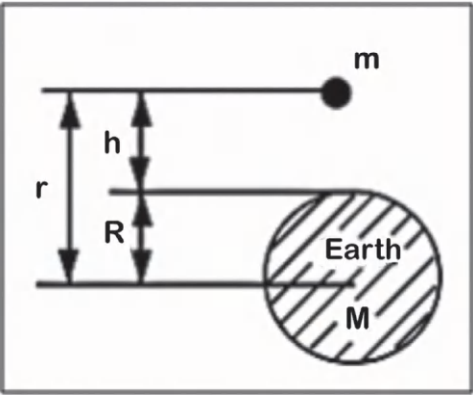


FIGURE 3.9A Variation in acceleration due to gravity due to altitude.

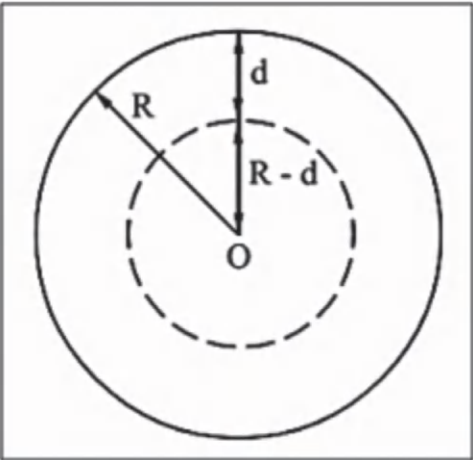


FIGURE 3.9B Variation in acceleration due to gravity due to depth.



FIGURE 3.9C Variation of g with altitude and depth.

where

$$M = \frac{4}{3} \pi (R-d)^3 \rho$$

$$F = \frac{\frac{4}{3} G \pi (R-d)^3 \rho m}{(R-d)^2} = mg$$

or

$$g = \frac{4}{3} G \pi (R-d) \rho$$

Here, g is directly proportional to $(R-d)$, so g will be less and less inside the Earth as we go down from its surface. Then, g is zero at the centre of the Earth.

The following figure (Figure 3.9c; Wikipedia) shows the graphical representation of the acceleration due to gravity from place to place.

3.4 MEASUREMENT OF GRAVITY ON LAND

The mean value of gravity at the surface of the Earth is approximately 9.8 m/s^2 , or 980,000 mGal. It is observed that the Earth's rotation and flattening cause gravity to increase by roughly 5,186 mGal from equator to pole, which is a variation of only about 0.5%. Accordingly, measurements of gravity are of two types. The first corresponds to the determination of the absolute magnitude of gravity at any place; the second consists of measuring the change in gravity from one place to another. In recent years, geophysical exploration has become an important endeavour for dealing with several issues related to the exploration of the subsurface Earth, especially in gravity prospecting, where we measure the small changes in gravity caused by underground structures and density variability as accurately as possible. The reliability of geophysical prospecting using gravimetry requires an instrumental sensitivity of the order of 0.01 mGal. It is very difficult to design an instrument to measure the absolute value of gravity. High precision of the portable gravimeter is enough to be used easily in different places. In India, national regional gravity surveys are being carried out by the relative variations in the density contrasts determined with a gravimeter that may be converted to absolute values by calibration with absolute measurements made at selected stations.

3.4.1 ON THE EARTH'S SURFACE

3.4.1.1 Measurement of Absolute Gravity

3.4.1.1.1 Pendulum Method

The classical method of measuring gravity is with a pendulum. A simple pendulum consists of a heavy weight suspended at the end of a thin fibre. The compound (or reversible) pendulum, first described by Henry Kater in 1818, allows more exact measurements (Nettleton, 1939). Since gravity is an acceleration, its measurement should imply the determination of length (L) and time (T). Normally, a pendulum will suffice to measure the absolute value of g at a particular place. But often, the precision required in measuring T is not achievable. However, such measurement can be achieved only by classical pendulum or 'falling body' techniques. However, in gravity prospecting, it is the relative change in gravity (difference between two stations) that is determined. Absolute gravity values may be obtained by measuring the change in gravity from a known pendulum station. Such international gravity bases (IGB) for reference are published by the Survey of India (SOI), the Hawaiian Institute of Gravity, NGRI, and others. Modern instruments capable of rapid gravity measurements are called *gravimeter*.

The period of a physical pendulum may be written as

$$T = 2\pi \sqrt{\frac{I}{mgh}}$$

where I is the moment of inertia about the knife-edge, m is the total mass of the pendulum, h is the distance from knife-edge to the centre of gravity, and g is the acceleration due to gravity. In Kater's reversible pendulum, the factor $\frac{I}{mh}$ is replaced by L , the length of equivalent simple pendulum, as the former cannot be measured with high precision. The sensitivity of the compound pendulum is calculated in the following manner. We know

$$T = 2\pi \sqrt{\frac{I}{mgh}} \rightarrow \frac{T^2}{4\pi^2} = \frac{I}{mgh} \Rightarrow g = 4\pi^2 \frac{I}{T^2mh}$$

Differentiating,

$$\frac{dg}{dT} = 4\pi^2 \frac{I}{mh} \left(\frac{-2}{T^3} \right) = -\frac{2g}{T} \Rightarrow \frac{dg}{g} = -2 \frac{dT}{T}$$

The earlier study made using such apparatus was bulky but was used until the 1950s as the main method of making absolute gravity measurements.

3.4.1.1.2 Free Fall Method

Modern methods of determining acceleration due to gravity are based on observations of falling objects. For an object that falls from a starting position $\mathbf{z}_0$ with initial velocity u , the equation of motion gives the position z at time t as

$$z = z_0 + ut + \frac{1}{2}gt^2$$

The absolute value of gravity is obtained by fitting a quadratic to the record of position versus time. In an absolute measurement of gravity, a laser beam is split along two paths that form a Michelson interferometer (Nettleton, 1939). The setup of the equipment is decided in such a way that the horizontal path is made as a fixed length, while the vertical path is reflected off a corner-cube retroreflector, which is released at a known instant and falls freely. The path of free fall is about 0.5 m long. The cube falls in an evacuated chamber to minimize air resistance. A photo-multiplier and a counter permit the fringes within any time interval to be recorded and counted. The intensity of the recombined light fluctuates sinusoidally with increasing frequency the further and faster the cube falls. The distance between each zero crossing corresponds to half the wavelength of the laser light, and so the distance travelled by the falling cube in any time interval may be obtained. The times of the zero crossings must be measured with an accuracy of 0.1 ns (10^{-10} s) to give an accuracy of 1 μGal in the gravity measurement.

Despite the compactness of the apparatus and its not being portable enough for gravity surveying in the area, it gives measurements of the absolute value of gravity with an accuracy of about 0.005–0.010 mGal (5–10 μGal). A disadvantage of the free fall method is the resistance of the residual air molecules left in the evacuated chamber. This effect is reduced by placing the retroreflector in a chamber that falls simultaneously with the cube so that, in effect, the cube falls in still air. Air resistance is further reduced in the rise-and-fall method.

3.4.1.1.3 Rise-and-Fall Method

In the original version of the rise-and-fall method, a glass sphere was fired vertically upward and fell back along the same path (Nettleton, 1939). Timing devices at two different levels registered the times of passage of the ball on the upward and downward paths. In each timer, a light beam passed through a narrow slit. The distance h between the two-timing levels (around 1 m) was measured accurately by optical interferometry.

Let the time spent by the sphere above the first timing level be T_1 , and the time above the second level be ' T_2 '; further, let the distances from the zenith level to the timing levels be z_1 and z_2 , respectively. The corresponding times of fall are $t_1 = \frac{T_1}{2}$ and $t_2 = \frac{T_2}{2}$. Then,

$$z_1 = \frac{1}{2} g \left(\frac{T_1}{2} \right)^2$$

with a similar expression for the second timing level. Their separation is

$$h = z_1 - z_2 = \frac{1}{8} g (T_1^2 - T_2^2)$$

The following is a simple expression for how the value of gravity can be obtained:

$$g = \frac{8h}{(T_1^2 - T_2^2)}$$

This asymmetry helps minimize the effects of air resistance. In a modern variation, Michelson interferometry is used as in the free fall method.

3.4.1.2 Measurement of Relative Gravity

Straightforward application of Hooke's law led to the development of the first gravimeter, or a gravity meter. A mass m suspended from a spring of length s_0 causes it to stretch to a new length (s). It is clear that the extension, or change in length, of the spring is proportional to the restoring force of the spring that gives the value of the gravity according to

$$F = mg = -k(s - s_0) \quad (3.12)$$

where k is the elastic constant of the spring. The gravimeter is calibrated at a known location. If gravity is different at another location, the extension of the spring changes, and from this the change in gravity can be computed.

As mentioned earlier, this type of gravimeter, based directly on Hooke's law, is called a *stable type*. It has been replaced by more sensitive *unstable or astatic types*, which are constructed so that an additional force acts in the same direction as gravity and opposes the restoring force of the spring (Telford et al., 1990). The instrument is then in a state of unstable equilibrium. This condition is realized through the design of the spring. It is observed that if the natural length s_0 can be made as small as possible, ideally 0, then, in accordance to eq. 3.12, the restoring force is then proportional to the physical length of the spring instead of its extension. The *zero-length spring*, first introduced in the *LaCoste–Romberg gravimeter*, is now a common element in modern gravimeters. It has been reported that the spring is usually of the helical type. When a helical spring is stretched, the fibre of the spring is twisted; the total twist along the length of the fibre equals the extension of the spring as

a whole. It is interesting to note that during manufacture of a zero-length spring, the helical spring is given an extra twist, since zero-length spring has the tendency to uncoil. An increase in gravity stretches the spring against its restoring force, and the extension is augmented by the built-in pre-tension.

The operation of a gravimeter is illustrated in Nettleton (1939). A mass is supported by a horizontal rod to which a mirror is attached. The position of the rod is observed with a light beam reflected into a microscope (Nettleton, 1939). If gravity changes, the zero-length spring is extended or shortened and the position of the rod is altered, which deflects the light beam. The null-deflection principle is utilized. An adjusting screw changes the position of the upper attachment of the spring, which alters its tension and restores the rod to its original horizontal position as detected by the light beam and microscope. Most importantly, the turns of the adjusting screw are calibrated in units of the change in gravity, usually in mGal.

The gravimeter is light, robust, and portable. The levelling of the equipment is followed by a reading from the gravimeter at site. The gravimeter has a sensitivity of about 0.01 mGal (10 μ Gal), which shows high sensitivity of the measuring gravimeter and makes it possible to measure small changes in the subsurface properties at the site.

3.4.1.2.1 Sensitivity of the Unstable or Astatized Types of Gravimeter

The sensitivity of unstable or astatized types of gravimeter may be calculated in the following manner. Suppose a spring of length l has been stretched to $(l + \delta l)$ as a result of an increase of gravity δg , increasing the weight of the suspended mass m (Figure 3.10). The extension of the spring is proportional to the extending force (Hooke's law) (Nettleton, 1939). Thus:

$$m \delta g = k \delta l \Rightarrow \delta l = \frac{m}{k} \delta g \Rightarrow \frac{\delta l}{\delta g} = \frac{m}{k} \quad (3.13)$$

This equation is valid for static gravimeters, where k is the elastic constant of the spring, or spring constant. δl must be measured to a precision of 10^{-7} cm in instruments called gravimeters. If $\delta l = 1$, $k = m \delta g$. Thus, spring constant k is the weight required to produce unit elongation in the spring, which depends on the elastic constant of the material, the diameter of the wire of the spring. This system vibrates in simple harmonic motion (SHM).

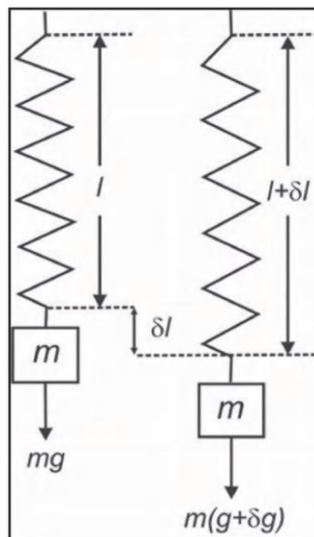


FIGURE 3.10 Extension (δl) of a spring due to additional gravitational pull (δg).

So the instrument can be made more and more sensitive by increasing m and reducing k . But a large mass and a weak spring make the instrument liable to collapse. On the other hand, if k is reduced to $(k - k')$ by the application of some force (astatizing force or labializing force), we get an astatic gravimeter. Then,

$$T = \frac{2\pi m}{k - k'} \quad (3.14)$$

where k' is the constant for labializing force. It is clear that the closer the value of k' approaches k , the greater will be the sensitivity.

The CG-5 is the latest gravity meter from Scintrex Ltd. The CG-5 can be operated with minimal operator training, and automated features significantly reduce the possibility of reading errors. The features of this gravimeter are: resolution mGal, 0.001 mGal (=1 μ Gal); standard deviation, <5 μ Gal; operating range, 8,000 mGal without re-setting; automated corrections, tide, instrument tilt, temperature, and rejection. Following is the photograph of a CG-5 Scintrex automated gravity meter (CG5.f0.manual.book).

3.4.1.3 Measurement of the Component of Gravity

Users of gravimeters should note that gravimeters effectively respond only to the vertical component of the gravity of an anomalous mass measured at the site. Consider the gravitational effect of an anomalous mass dg with horizontal and vertical components dg_x and dg_z , respectively, on the local gravity field g (Figure 3.11a). We can show that θ is usually insignificant since $\delta g_z \ll g$ (Radogna et al., 2004).

Solving the rectangle of forces:

$$g + dg = [(g + dg_z)^2 + dg_x^2]^{1/2} = [g^2 + 2gdg_z + dg_z^2 + dg_x^2]^{1/2}.$$

Terms in d^2 are insignificantly small and can thus be ignored. Binomial expansion then gives

$$\begin{aligned} g + dg &= g + dg_z \\ dg &= dg_z \end{aligned}$$

Therefore, measured perturbations in gravity effectively correspond to the vertical component of the attraction of the causative sources.

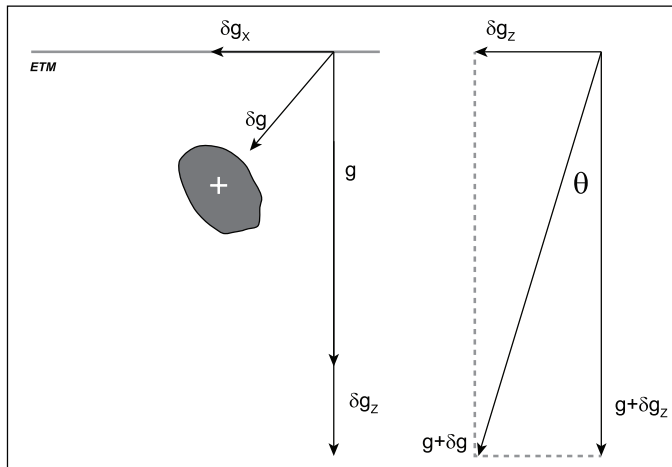


FIGURE 3.11A Gravity measures only vertical component.

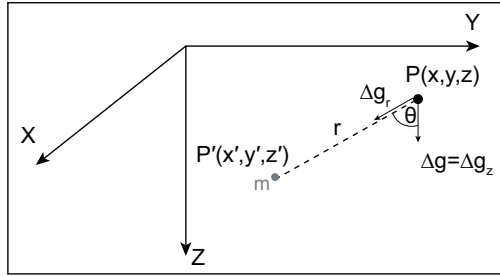


FIGURE 3.11B Schematic diagram of point mass.

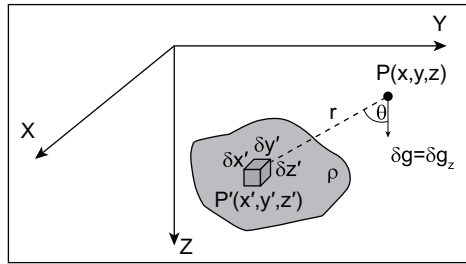


FIGURE 3.11C Schematic diagram of irregular shape.

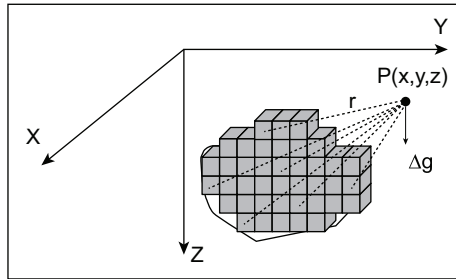


FIGURE 3.11D Schematic diagram of irregular shape for the whole body.

3.4.1.3.1 Gravity Anomaly Point Mass (Figure 3.11b)

From Newton's law, we get

$$\Delta g_r = \frac{Gm}{r^2}$$

$$\text{We know } \Delta g = \Delta g_z = \frac{Gm}{r^2} \cos\theta = \frac{Gm}{r^2} \frac{(z' - z)}{r} = \frac{Gm(z' - z)}{r^3}.$$

3.4.1.3.2 Gravity Anomaly: Irregular Shape (Figure 3.11c)

We have

$$\Delta g = \frac{Gm(z' - z)}{r^3}$$

For $\delta m = \rho \delta x' \delta y' \delta z'$, we derive

$$\delta g = \frac{G\rho(z' - z)\delta x'\delta y'\delta z'}{r^3}$$

with ρ the density in gm/cc, and $r = \sqrt{(x' - x)^2 + (y' - y)^2 + (z' - z)^2}$.

For the whole body (Figure 3.11d):

$$\Delta g = \sum \sum \sum \frac{G\rho(z' - z)}{r^3} \delta x' \delta y' \delta z'$$

If $\delta x' \delta y' \delta z'$ approaches 0, then

$$\Delta g = \iiint \frac{G\rho(z' - z)}{r^3} dx' dy' dz'$$

The gravitational anomaly can be efficiently computed. The direct problem in gravity is straightforward: Δg is found by summing the effects of all elements which make up the body.

3.4.2 IN BOREHOLES

It is reported that gravity was first measured in boreholes in the 1960s. In these years, Esso and the US Geological Survey (USGS)/LaCoste–Romberg gravimeter types were in use for measuring gravity in boreholes. They have sensitivities of ~ 0.01 mGal. Care is taken for temperature control because of the geothermal gradient. Gravimeters must also allow for deviations from the vertical of boreholes. It is reported that the USGS/LaCoste–Romberg meter can be levelled up to 6.5° off vertical and is kept at 101°C by a thermostat. Thus, it will not work at temperatures higher than this. It takes ~ 5 minutes to make a reading.

These measurements are important for determining densities. In recent years, it has been established that borehole gravimeters are the best borehole density loggers in existence. They are sufficiently sensitive to monitor reservoir depletion as water replaces oil. The density in the interval between the measurements is calculated in the following manner.

Let g_1 and g_2 be the values of gravity measured in a vertical borehole at heights h_1 and , respectively, above the reference ellipsoid (Figure 3.12). The difference between g_1 and g_2 is due to the different heights and to the material between the two measurement levels in the borehole (Telford et al., 1990). The value of g_2 will be larger than the value of g_1 , for two reasons. First, because the lower measurement level is closer to the Earth's centre, g_2 will be greater than g_1 by the amount of the combined elevation correction, namely, $[0.3086 - (0.0419\rho \times 10^{-3})] \Delta h$ mGal, where $\Delta h = h_1 - h_2$. It is noteworthy that the second, at the lower level h_2 , the gravimeter experiences an upward Bouguer attraction due to the materials between the two measurement levels. This reduces the measured gravity at h_2 and requires a compensating increase to g_2 of the amount $(0.0419\rho \times 10^{-3}) h$ mGal. The difference Δg between the *corrected* values of g_1 and g_2 after reduction to the level h_2 is then

$$\begin{aligned} \Delta g &= (0.3086 - 0.0419\rho \times 10^{-3}) \Delta h - 0.0419\rho \times 10^{-3} \Delta h \\ &= (0.3086 - 0.0838\rho) \Delta h \end{aligned}$$

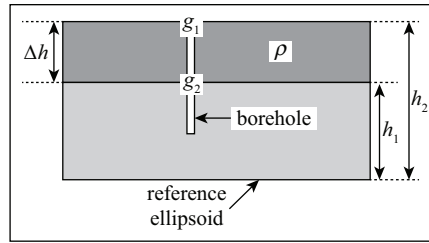


FIGURE 3.12 Geometry for computation of the density of a rock layer from gravity measurements made in a vertical borehole.

Rearranging this equation gives the density ρ of the material between the measurement levels in the borehole:

$$\rho = (3.683 - 11.93 \frac{\Delta g}{\Delta h}) \cdot 10^3 \text{ kg m}^{-3}$$

Recent measurements of the gravity in boreholes demonstrated that if borehole gravity measurements are made with an accuracy of ± 0.01 mGal at a separation of about 10 m, the density of the material near the borehole can be determined with an accuracy of about $\pm 10 \text{ kg m}^{-3}$. More than 90% of the variation in gravity in the borehole is due to material within a radius of about $5 \Delta h$ from the borehole (about 50 m for a distance $\Delta h \approx 10$ m between measurement levels). This is much larger than the lateral range penetrated by gamma–gamma logging.

3.5 MEASUREMENT OF GRAVITY ON MOVING PLATFORMS

3.5.1 SEA SURVEYS

It is reported that measurement of gravity at sea was first done by lowering the operator and the instrument in a diving bell. This is no longer done for shipborne marine gravimetry, because it is slow and expensive. Now, two methods are used.

3.5.1.1 Lowering the Meter onto the Seafloor (~0.1 mGal Accuracy)

The meter is operated by remote control. In marine surveys, Gulf and LaCoste–Romberg gravimeters are adapted for the purpose. Errors arise due to wave motion at the surface, of which decrease with depth is an important constraint. It is advisable to install the instrument on rock and not into the mud. It is necessary to measure the water depth for which a pressure gauge is deployed that gives a read-out on the same panel as the gravity reading. This method is used to study gravity anomalies of small extent, for example, salt domes under the sea. The sensitivity of these gravimeters is ~ 0.1 mGal. It is a very expensive marine geophysical survey using shipborne gravity meter, as the ship has to stop for each reading, so it is time-consuming.

3.5.1.2 The Meter On Board Ship (Recently Improved from ~2 to 0.2 mGal Accuracy)

This is fundamentally difficult because the ship experiences accelerations up to 10% of g (100,000 mGal). The horizontal motions are compensated for by mounting the meter on a gyroscopically controlled stable platform. The vertical motions are compensated for by averaging over a long period and by damping the meter heavily, for example, by using a meter with a 5-minute natural period. This results in long period anomalies only being measured, that is, a heavily damped meter

functions as a low-pass filter. It is a proven fact that the accuracy of shipborne gravimetry depends on the state of the sea, and the measuring devices for gravity by Gulf, LaCoste–Romberg, Bell, and Askania are available for such marine gravity surveys.

3.5.2 AIR SURVEYS (ACCURACIES ~1–5 mGal)

A series of research studies on airborne gravimetry showed that several problems due to the acceleration of the aircraft have not yet been completely solved. However, a significant breakthrough has been made with the advent of the Global Positioning System (GPS). Reasonably, a good regional survey has been achieved, where accuracies of a few mGal have been reported. Airborne gravity surveying has the potential to greatly reduce the expense of gravity surveys by enhancing coverage of the surveyed areas, but how to use the results is still reportedly controversial. It is observed that airborne gravity-based observations have no corroboration with land results because of high discrepancies in airborne data which are beyond the tolerable limit and much larger than the ‘official’ errors, which in turn suggests that the true accuracy of these surveys is worse than the calculated precision, a common situation in science.

3.5.3 SPACE MEASUREMENTS

Determining the gravity field of the Earth from space involves measuring the height of a satellite above sea level by radar altimetry. A series of satellites have been used, including Skylab, GEOS3, Seasat, Geosat, ERS1, and ERS2. The Seasat was launched in 1978 into a circular orbit with an altitude of 800 km. It circled the Earth 14 times each day and covered 95% of the Earth’s surface every 36 hours (Foulger and Pierce, 2014). The position of Seasat was continually tracked by laser sites whose coordinates with respect to the spheroid are known, and the position of the Seasat was measured three-dimensionally. The satellite continually transmitted a radar signal which bounced off the sea surface. The two-way travel time was measured.

- H^* was derived from tracking.
- The satellite measures h .
- H_g , the geoid height, was calculated.

The ‘footprint’ of the radar beam on the surface of the sea was 2–12 km wide, and this represents the diameter of the ‘points’ that were measured. The precision of measurement was 10–20 cm. The gravity fields returned were used to study variations in the Earth’s mass and density distribution, since these are related directly to geoid topography.

It is observed that recent modern advances in gravimeters include the additional feature of direct digital read-outs, which speed up measurements, and the use of GPS navigation in the case of moving platforms has become a routine requirement for marine surveys. This has greatly improved the accuracy of the Eötvös correction, reducing the error from this source from ~ 2 mGal to 0.2 mGal. Interestingly, the advent of information technology brought a revolution in creating a data repository for reasonable gravity fields on regional scales, which are now available for most of the Earth via the internet, so it is becoming less important for oil companies to do their own geophysical surveys of the targeted region. A discussion of the comparative accuracies of various survey methods may be found in detail in Foulger and Pierce (2014).

3.6 GRAVITY SURVEYING

A gravity survey may be conducted over a profile or over an area, depending on the need. The station spacing may vary from a few meters (detailed mineral or geotechnical surveys) to a few

kilometres in regional reconnaissance surveys. The following factors must be considered in designing a survey:

- The survey must include at least one station where absolute g is known.
- The station spacing must be carefully selected so that it fits the anomaly scale.
- The elevations of all stations must be known or measured to ± 10 mm. Uncertainties in the elevation of the gravity stations account for the greatest errors in reduced gravity values on land. Levelling work associated with gravity survey is to be carried out with desirable accuracies, preferably with a Differential Global Positioning System (DGPS) instrument, a state-of-the-art DGPS system which has the capacity to measure both position location and elevation within a short period of observation time. A field photo for data acquisition system is incorporated for easy understanding (GSI FSP 2018–2019, code: M1AGS-GPM/NC/SR/2018/17238).
- Latitudes must be known to ± 10 m from a toposheet.
- The selection of the observation stations must be at lesser undulation or stable topography to maintain accuracy in measurements.
- Access is important, which often means keeping stations to existing roads or waterways if there are no roads.
- In the design of the gravity survey, station spacing and accuracy are to be judiciously decided at the time of designing the gravity survey.

It is important to realize that no amount of computer processing can compensate for poor experiment design. Linear features may be studied using one or more profiles, two-dimensional features may require several profiles plus some regional points, and for some special objectives, such as in the determination of the total subsurface mass, widely spaced points over a large area may be appropriate.

The following field procedure is usually adopted:

- Measure a base station.
- Measure more stations.
- Re-measure the base station approximately every two hours.

If the survey area is large, time can be saved by establishing a conveniently sited base station to reduce driving. This is done as follows:

Measure: base 1 $\rightarrow$ new base station $\rightarrow$ base 1 $\rightarrow$ new base station $\rightarrow$ base 1

It is suggested that results, as mentioned, be in three estimates, and the difference in gravity between base 1 and the new base station is recorded; from this, gravity at the new base station may be calculated. The new base station can then be re-measured at two-hourly intervals instead of the base 1. This method is known as looping and is done to carry out drift corrections, which depend on the tidal forces on the day of the survey and the elastic parameter of the spring of the gravity meter, which is a right procedure to establish an absolute base station within the survey area. The process of obtaining the relative difference between the established base International Gravity Standardization Net 1971 (IGSN, 1971, prepared by Morelli et al., 1971, for the Air Force Cambridge Research Laboratories, European Office of Aerospace Research and Development) and the field base is carried out in a systematic way:

- A minimum of three sets of readings at the field base and the IGSN base is observed and recorded.
- The average difference is taken to obtain the value of the field base. An example is given in Table 3.2.

Readings from S. no. 1 to 3, 3 to 5, and 5 to 7 form three sets of readings for the base tie.

TABLE 3.2**Gravity Base Tie**

S. No.	Station	Time	Reading	Difference in mGal	Drift in mGal	Corrected Difference
1	Base 1	07:30	736.5			
2	Base 2 (IGSN- 71)	08:00	740.5	4.0	−0.0075	3.9925
3	Base 1	08:30	736.515	0.015	−0.015	
4	Base 2	09:00	740.525	4.01	−0.005	4.005
5	Base 1	09:30	736.525	0.01	−0.01	
6	Base 2	10:00	740.540	4.015	−0.025	3.99
7	Base 1	10:30	736.55	0.05	−0.05	3.995 is the average difference value of the two bases in mGal.

TABLE 3.3**Gravity Data Record Sheet**

Instrument No. and Make

Area with Toposheet No.:

Month/Year:

State:

Date:

Reading Time:

Traverse Id:

Name of the Observer:

District:

S. No.	Base/Station ID	Time	Reading (mGal)	Longitude (Deg)	Latitude (Deg)	Geology	Remarks	Landmarks

Besides recording the gravity data in the instrument, it is good practice to keep a record of data in the field notebook as well. The standard format for data recoding in the field notebook is provided in Table 3.3.

Transport during a gravity survey may be motor vehicle, helicopter, air, boat (in marshes), pack animal, or by walking. In very rugged terrain, geodetic surveying to obtain the station heights may be a problem.

3.6.1 REDUCTION OF GRAVITY DATA

Before the results of the gravity survey can be interpreted, it is necessary to correct for all variations in the Earth's gravitational field which do not result from the difference of density in the underlying rocks. This is a very important process, known as *gravity reduction or reduction to the geoid*, as sea level is usually the most convenient datum level. These corrections are for the drift, latitude, free air, Bouguer, and terrain or topographic correction applicable when the terrain is rugged. Eötvös correction is applicable when the data was collected from a moving platform, like a ship or an aircraft. Isostatic correction is required when gravity data is used for deep crustal studies.

3.6.1.1 Drift Correction

Plotted is a graph of measurements made at the base station throughout the day. Drift may be non-linear, but it has to be assumed that it is linear between tie backs for most surveys. The drift correction incorporates the effects of instrument drift, uncompensated temperature effects, solid Earth and sea tides, and the gravitational attraction of the sun and the moon (GSI FSP 2018–2019, code: M1AGS-GPM/NC/SR/2018/17238).

3.6.1.2 Tidal Correction

This correction is based on the tidal effect on the Earth due to the attraction of the sun and the moon (Radogna et al., 2004). During a full moon, this may reach up to 0.3 mGal. But careful application of drift correction will take care of most of the tidal effects also. After drift and tidal corrections, g_{obs} can be computed using Δg , the calibration factor of the gravimeter, and the value of gravity at the base (Mickus, 1980).

3.6.1.3 Latitude Correction

It is a proven fact that gravity varies with latitude (the *latitude* of a point is the angle φ between the equatorial plane and the line joining that point to the centre of the Earth; the latitude of the equator is 0° , and that of the poles is 90°) because of the non-spherical shape of the Earth and because the angular velocity at a point on the Earth surface decreases from a maximum at the equator to 0 at the poles (Wikipedia; kullabs.com). The centripetal acceleration generated by this rotation has a negative radial component causing gravity to be reduced from the poles to the equator. Again, the equatorial radius is 21 km more than the polar radius, resulting in an increase of gravity value from the equator to the poles by some 5,186 mGal. An alternative approach (the variation of the gravity value with latitude is given by the equation [Moritz, 1980; IAG, 1980]):

$$g_\varphi = 978.0327(1 + 0.0052790414 \sin^2\varphi - 0.0000232718 \sin^4\varphi + 0.0000001262 \sin^6\varphi) \text{ Gal} \quad (3.15)$$

The value of g_φ gives the predicted value of gravity at sea level at any point on the Earth's surface and is subtracted from the observed gravity to correct for latitude variation. International gravity formulae 1930 and 1980 are in use. As at present, the gravity data is reduced only with respect to the 1980 formula; for reducing the earlier data of 1930 datum, the following equation can be used. The entirety of the international gravity formulae is given in Chapter 16 of this book. The actual difference between the 1930 international gravity formula and the 1980 international gravity formula is:

$$(-16.3 + 13.7 \sin^2\varphi) \text{ mGal, where } \varphi \text{ is the latitude of the base} \quad (3.16)$$

The theoretical gravity value corresponding to the latitude of a particular point as obtained from the preceding formula (eq. 3.15) is removed from the observed gravity, also referred to as station gravity. In regional surveys, where the base is generally tied to a known gravity station and the anomalies are calculated by removing the theoretical gravity from the absolute observed gravity, no latitude correction will be necessary. When the surveys are of a detailed nature, a correction of $0.8122 \sin 2\varphi$ mGal/N-S km will be applicable, where φ is the mean latitude of the area of survey.

3.6.1.4 Elevation Correction

Corrections for the differing elevations are made in three parts, which are described for conceptual background (Saibi, 2018):

- Free air correction
- Bouguer correction
- Terrain correction

3.6.1.4.1 Free Air Correction

Corrections for the decrease in gravity with height (in free air) resulting from the increased distance from the centre of the Earth (Figure 3.13). To reduce to the datum of an observation taken at a height h in metre from the mean sea level:

$$\text{Free air correction (FAC)} = 0.3086 h \text{ mGal/m} \quad (3.17)$$

It is evident that FAC is positive for an observation point above the datum to correct for the decrease in gravity with observation, whilst FAC is negative for an observation point below the datum (usually for underwater measurements, surveys in mines, etc.).

It solely takes into account the variation of gravity with distance from the centre of the Earth and does not correct for the gravitational effect of the rocks present between the observation point and the datum.

3.6.1.4.2 Bouguer Correction

Removes the effect of the rock mass by approximating an infinite horizontal slab with thickness equal to the elevation of the observation point above the datum (Figure 3.14). If ρ is the density of the rock, then

$$\text{Bouguer correction (BC)} = 2\pi G \rho h = 0.04191 \rho h \text{ mGal/m} \quad (3.18)$$

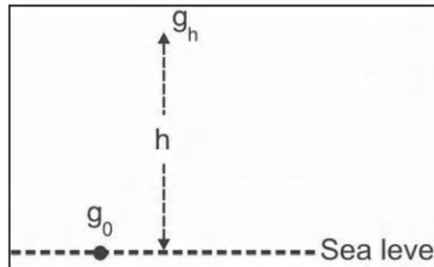


FIGURE 3.13 Free air effect.

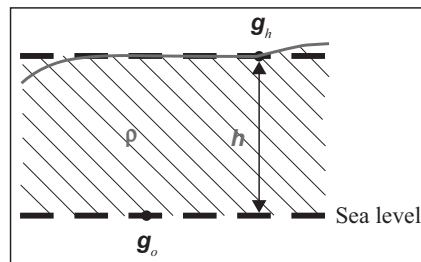


FIGURE 3.14 Bouguer effect.

where ρ is in gm/cc. It is now clear that, on land, Bouguer correction must be subtracted, as the attraction due to the mass of the rock between the datum and the observation point must be removed from the observed gravity. In sea,

$$BC = 2\pi G (\rho_r - \rho_w) z$$

Where:

ρ_r = density of rock in land

ρ_w = density of seawater

z = depth of the sea

If the station is at a lower level than the reference station, its gravity value is decreased because of the lack of attraction of the absent material between it and the reference level. Therefore, the Bouguer correction and the free air correction are always of opposite sign. Usually, free air and Bouguer corrections are combined together, and the total correction is called the combined elevation correction.

Therefore:

$$\begin{aligned} \text{A combined elevation correction factor} &= (0.3086 - 0.04191 \times 2.67) \text{ h mGal/m} \\ &= 0.1967 \text{ h mGal/m} \end{aligned} \quad (3.19)$$

This correction is positive for stations higher than the datum, and negative for stations lower than the datum.

3.6.1.4.3 Terrain Correction

Bouguer correction assumes a subdued topography. Additional terrain corrections must be applied in rugged areas where measurements are near *mountains* or *valleys* (Saibi, 2018). It is documented fact that if the station is next to a mountain, there is an upward force on the gravimeter from the mountain that reduces the reading, which was first noticed in topographic surveys run near the Himalayan mountains. Conversely, if the gravity station is next to a valley, there is an absence of the downward force on the gravimeter assumed in Bouguer correction, which reduces free air anomaly. In both cases, terrain correction is added to the Bouguer anomaly.

We can show that the terrain correction (Δg_T) for a hill adjacent to a gravity station is computed by dividing the hill into a number of vertical prisms (Lowrie, 2007). The contribution of each vertical element to the vertical acceleration at the point of observation P is calculated by assuming cylindrical symmetry about P . The height of the prism is h ; its inner and outer radii are r_1 and r_2 , respectively; the angle subtended at P is φ_0 ; and the density of the hill is ρ (Lowrie, 2007). Let the sides of a small cylindrical element be dr , dz , and $rd\varphi$; its mass is $dm = \rho r dr dz d\varphi$; and its contribution to the upward acceleration caused by the prism at P is

$$\Delta g = G \frac{dm}{r^2 + z^2} \cos\theta = G \frac{\rho r dr dz d\varphi}{r^2 + z^2} \frac{z}{\sqrt{r^2 + z^2}}$$

Combine and rearrange the terms and the order of integration gives the upward acceleration at P due to the cylindrical prism:

$$\Delta g_T = G\rho \int_{\varphi=0}^{\varphi_0} d\varphi \int_{r=r_1}^{r_2} dr \left(\int_{z=0}^h \frac{z dz}{(r^2 + z^2)^{\frac{3}{2}}} \right) r dr$$

The integration over φ gives φ_0 ; after further integration over z , we get

$$\Delta g_T = G\rho\varphi_0 \int_{r=r_1}^{r_2} \left(\frac{r}{\sqrt{r^2 + h^2}} - 1 \right) dr$$

Integration over r gives the upward acceleration produced at P by the cylinder:

$$\Delta g_T = G\rho\varphi_0 \left[\left(\sqrt{r^2 + h^2} - r_1 \right) - \left(\sqrt{r^2 + h^2} - r_2 \right) \right]$$

The direction of Δg_T , as explained by Lowrie (2007), is upward, opposite to gravity; the corresponding terrain correction must be added to the measured gravity.

There is a procedural discussion for conducting terrain corrections, which are done by hand using a transparent graticule, or by a computer if a digital terrain map is available. The graticule is placed on a map, and the average height of each compartment is estimated (Hammer, 1939). A ‘Hammer chart’ is then used to obtain the correction (Hammer, 1939). The prepared chart gives the correction for a particular distance from the station, which is generally worked out assuming a block of constant height for each compartment. Other charts are available, for example, the Sandberg tables, which provide for larger terrain differences and assume a sloping terrain. Recently, terrain corrections have now been done with digital terrain maps, and several computational algorithms have been developed, as doing the work by hand is very time-consuming and involves a lot of repetition.

3.6.1.5 Eötvös Correction

A ship or an aircraft moving towards east or west has its centrifugal force increased or decreased, thus changing the gravity. The Eötvös correction is given by the following formula:

$$E_c = 7.503 V \cos\varphi \sin\alpha + 0.004154 V^2$$

where V is in knots, φ is the latitude, and α is the direction of movement w.r.t. the north.

3.6.1.6 Isostatic Correction

This is similar to terrain correction but based on tables prepared on the hypothesis of isostasy. This correction has no relevance in gravity prospecting.

3.7 GRAVITY ANOMALIES

The *gravity anomaly* at a point may be defined as the difference between the observed gravity and the theoretically computed one for the point using the international gravity formula:

$$\delta g = \delta g_{\text{obs}} - g_\varphi \quad (3.20)$$

To make the observed and theoretical gravity values comparable, the reduction steps are adopted, each step of reduction resulting in a particular type of anomaly.

3.7.1 FREE AIR ANOMALY

The anomaly mainly represents the topographic features superposed over the effects of the subsurface. Therefore:

$$\text{FAA} = \text{observed gravity} + \text{free air correction} - \text{theoretical gravity} = g_{\text{obs}} + \text{FAC} - g_\varphi$$

3.7.2 BOUGUER ANOMALY

It is defined as:

$$BA = g_{\text{obs}} + FAC (-, +) BC + TC - g_{\phi} \quad (3.21)$$

As the Bouguer anomaly is free from the effect of any material lying between the observation point and the datum, it represents the variation in the sub-datum mass distribution. Hence, a zero BA indicates mass distribution within the Earth perfectly coinciding with that of the ideal spherical Earth. But the continental elevated portions usually show negative BA, revealing lower-than-normal density of rock in these regions. Similarly, the oceanic regions usually have a positive BA, suggesting higher-than-normal density for the rocks there in the oceanic crust. The contributions arising due to the lateral inhomogeneity are called the BA, or simply gravity anomaly.

3.7.3 ISOSTATIC ANOMALY

These are the anomalies obtained after correcting the BA for the isostatic behaviour of the crust. This correction is made by taking into account the surface topographic mass distribution. Therefore:

$$IA = g_{\text{obs}} + FAC - BC + TC + \text{isostatic correction} - g_{\varphi} \quad (3.22)$$

A zero isostatic anomaly indicates complete compensation, while positive anomalies reveal undercompensation, and the negative anomaly overcompensation. The isostatic anomalies are of little significance in prospecting, except in locating possible zones of strain accumulation and tectonic activity.

3.7.4 DATA FORMAT FOR REDUCTION OF GRAVITY DATA

The data format for a simple software application that corrects for free air, Bouguer slab, and normal variation of gravity is given in Table 3.4.

Terrain correction/topographic correction is a very cumbersome process. This correction is normally ignored, unless it is very essential, where the anomaly sought (signal) is of the same order as that of the topographic noise. Descriptions of the terrain conditions at the point of observation may be noted, as it is required during terrain corrections.

TABLE 3.4

Data Format for Reduction of Gravity Data

S. No.	Stn. Id.	Long. (Degree)	Lat. (Degree)	Elev. (Ellip.)* m	Elev. (msl) m	Stn. Grav. (1971) mGal	Theo. Grav. (1980) mGal	Boug. Anom. mGal	Free Air Anom. mGal	Remarks/Process of Converting Ellipsoid to msl

Note: *If DGPS has been used for elevation.

3.8 ISOSTASY

Isostasy is the study of load compensation for a given depth; for example, the mountain belts on the Earth's surface are found to be compensated for at depth. The difference in the apparent angle of the star with the plumb line (i.e. the difference between astronomic and geodetic) between two stations, Kaliana (in the foothills of the Himalayas) and Kalianpur (on the Indo-Ganges plain), in North India has led to the theory of isostasy. There are two main theories of isostasy, one due to G.B. Airy, who visualized a lower density (roots) under mountains and a higher density (anti-roots) under the oceans, and the other due to J.H. Pratt, who suggested that crustal thickness is uniform and that the mountains and oceans are underlain by less-than-normal and more-than-normal densities.

3.8.1 THE AIRY–HEISKANEN MODEL

The Airy–Heiskanen model assumes local isostatic compensation, that is, the root zone of a mountain lies directly under it. The vertical column for the mountain of height h_1 in Figure 3.15a contains only crustal rocks of density ρ_c . The pressure at CC' due to the mountain, 'normal' crust of thickness (t), and root zone of thickness r_1 amounts to $(h_1 + t + r_1)\rho_c$. The vertical column below the 'normal' crust contains a thickness (t) of crustal rocks and thickness r_1 of mantle rocks; it exerts a pressure of $(t\rho_c + r_1\rho_m)$. For hydrostatic equilibrium, the pressures are equal. Equating, and noting that each expression contains the term $t\rho_c$, we get

$$r_1 = \frac{\rho_c}{\rho_m - \rho_c} h_1 \quad (3.23)$$

with a similar expression for the root of depth r_2 under the hill of height h_2 . We can determine the thickness r_0 of the anti-root of the oceanic crust under an ocean basin of water depth (d), and density ρ_w is given by

$$r_0 = \frac{\rho_c - \rho_w}{\rho_m - \rho_c} d \quad (3.24)$$

The Airy–Heiskanen model assumes an upper layer of constant density floating on a denser substratum. It has root zones of variable thickness proportional to the overlying topography. We compared with the seismological evidence and found that this scenario agrees broadly with seismic evidence for the thickness of the Earth's crust and reveals that the continental crust is much thicker than the oceanic crust. Its thickness is very variable, being largest below mountain chains, although the greatest thickness is not always under the highest topography, since Airy-type compensation suggests hydrostatic balance between the crust and the mantle.

3.8.2 THE PRATT–HAYFORD MODEL

Several studies on isostasy revealed that the Pratt–Hayford isostatic model incorporates an outer layer of the Earth that rests on a weak magmatic substratum. Differential expansion of the material in the vertical columns of the outer layer accounts for the surface topography, so that the higher the column above a common base, the lower the mean density of rocks in it. The vertical columns have constant density from the surface to their base at depth D below sea level (Figure 3.15b; Wikipedia).

If the rock beneath a mountain of height h_i ($i = 1, 2 \dots$) has density ρ_i , the pressure at CC' is $\rho_i(h_i + D)$. Beneath a continental region at sea level, the pressure of the rock column of density ρ_c is $\rho_c D$. Under an ocean basin, the pressure at CC' is due to water of depth d and density ρ_w on top of

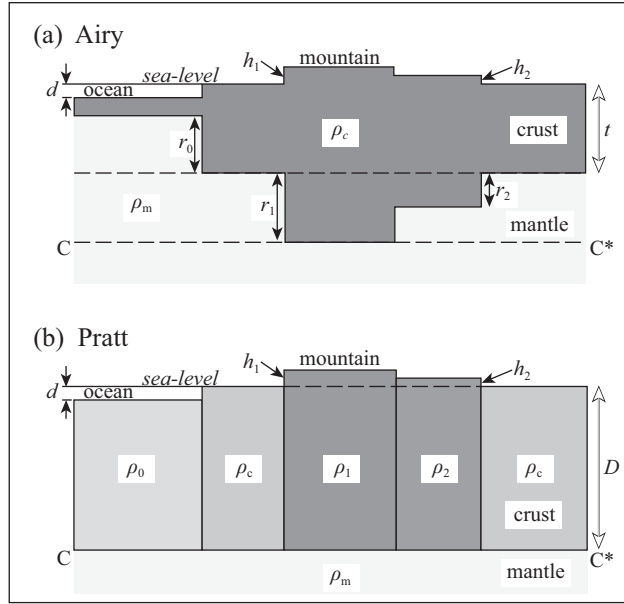


FIGURE 3.15 Local isostatic compensation according to (a) the Airy–Heiskanen model and (b) the Pratt–Hayford model.

Source: Wikipedia.

a rock column of thickness $(D - d)$ and density ρ_0 ; it is equal to $\rho_w d + \rho_0 (D - d)$. Equating these pressures, we get

$$\rho_i = \frac{D}{h_i + D} \rho_c \quad (3.25)$$

For the density below a topographic elevation h_i ,

$$\rho_0 = \frac{\rho_c D - \rho_w d}{D - d} \quad (3.26)$$

It is estimated that for the density under an oceanic basin of depth d , the compensation depth D is about 100 km. The Pratt–Hayford and Airy–Heiskanen models represent local isostatic compensation, in which each column exerts an equal pressure at the compensation level.

3.8.3 ISOSTATIC COMPENSATION AND VERTICAL CRUSTAL MOVEMENTS

We considered a case for the isostatic compensation model for both the Pratt–Hayford and Airy–Heiskanen models: the lighter crust floats freely on the denser mantle. The system is in hydrostatic equilibrium, and local isostatic compensation is a simple application of Archimedes' principle. A 'normal' crustal thickness for sea level coastal regions is assumed (usually 30–35 km), and the additional depths of the root zones below this level are exactly proportional to the elevations of the topography above sea level. The topography is then completely compensated (Lowrie, 2007). However, isostatic compensation is often incomplete. The geodynamic imbalance leads to vertical crustal movements.

It is very much understandable that the mountains are subject to erosion, which can disturb isostatic compensation. If the eroded mountains are no longer high enough to justify their deep root zones, the topography is isostatically overcompensated (Lowrie, 2007). It is also observed that buoyancy forces are created similar to a wooden block when floating in water; when pressed downward, the underwater part becomes too large in proportion to the amount above the surface. If the finger pressure is removed, the block rebounds in order to restore hydrostatic equilibrium. Similarly, the buoyancy forces that result from overcompensation of mountainous topography cause vertical uplift. The opposite scenario is also possible. If it is found that the topography has roots that are too small, the topography is isostatically undercompensated (Lowrie, 2007). This process is similar to when tectonic forces thrust crustal blocks on top of each other. Hydrostatic equilibrium is now achieved by subsidence of the uplifted region.

3.9 ROCK DENSITIES

The interpreter of gravity data aims to determine the subsurface variations of mass, and this process requires that the density of the material of interest, or the density contrast between the material of interest and the surrounding material, be known to assess the subsurface conditions. The density can be determined in many ways, with the best technique being acquiring rock samples within the study area and determining their average density.

It is a proven approach of determining the appropriate density to use in a gravity study by making a representative collection of rock samples with the aid of a geological map, whilst the specific gravity of a sample may be found directly by weighing it, first in air, and then in water, and applying Archimedes' principle of the fundamental physics. This gives its density ρ_r relative to that of water:

$$\rho_r = \frac{W_a}{W_a - W_w}$$

However, estimates of the densities for different rock types by this method show a large amount of discrepancies due to large scatter about their means, and the ranges of values for different rock types overlap (Dobrin, 1976). The densities of igneous and metamorphic rocks are generally higher than those of sedimentary rocks. This method is adequate for the reconnaissance of an area. Unfortunately, it is often difficult to ensure that the surface collection of rocks is representative of the rock types in subsurface structures, so alternative methods of determining the appropriate density are usually employed.

During the ground gravity survey boom in the 1930s, the question of estimating the appropriate correction density (as we call it now) or reduction density (as it was commonly called in the past) arose. This quantity is necessary for processing gravimetric measurements into the form of complete Bouguer anomalies. Traditional measurements on rock samples gradually turned out to be rather problematic, and so the gravimetric method was proposed (Nettleton, 1939). It is important to mention that, alternately, density can be measured in vertical boreholes drilled to explore the nature of a presumed structure, which is very useful, once the density in the borehole has been determined, in refining the interpretation of the structure. Most commonly for reliable and quick information, density can also be estimated from experimental relationships relating to compressional seismic velocities (obtained from seismic refraction surveys) and density (Nafe and Drake, 1957; Birch, 1961), which help the interpreter to use average density values from tables obtained from the measurements of numerous rocks, soil, and mineral samples (Johnson and Olhoeft, 1984; Telford et al., 1990). Table 3.5 shows the density range for the common sediments and sedimentary rocks usually encountered in environmental and engineering surveys (Nettleton, 1939).

Nettleton's method involves conducting a gravity survey over a topographic feature and reducing the data using a suite of densities by identifying results in an anomaly that correlates the least with

TABLE 3.5**Density Range of Common Sediments and Sedimentary Rocks.**

Rock Type	Density (gm/cm³)
Soil	1.20–2.40
Gravel	1.70–2.40
Alluvium	1.96–2.00
Chalk	1.53–2.60
Sand	1.70–2.30
Sandstone	1.60–2.75
Silt	1.80–2.20
Shale	1.75–3.20
Gypsum	2.20–2.60
Limestone	1.93–2.90

the topography, which is advantageous in the sense that bulk density is determined irrespective of the volume of a small sample (Zahorec et al., 2023). If the variation in gravity over the area is small, we may rearrange the Bouguer equation:

$$BA = BA_{\text{ave}} + \delta BA$$

where BA is ‘Bouguer anomaly’ at station, BA_{ave} is the average BA over the whole area, and δBA is the small increment of BA . The standard Bouguer anomaly equation is

$$BA = g_{\text{obs}} - g_{\varphi} + FAC - BC + TC$$

Thus:

$$g_{\text{obs}} - g_{\varphi} + FAC = BA_{\text{ave}} + \delta BA + BC - TC = \rho \left(0.04191h - \frac{TC}{2000} \right) + BA_{\text{ave}} + \delta BA$$

For Hammer charts using $\rho = 2,000 \text{ kg/m}^3$, this is an equation of the form $y = mx + c$ if δBA is small. If the line is plotted:

$$g_{\text{obs}} - g_{\varphi} + FAC: \left(0.04191h - \frac{TC}{2000} \right)$$

The form of scatter about a straight line provides a conspicuous distribution of the data. A line can be fitted to this using least squares, and this will have a gradient of ρ (Figure 3.16) (Nettleton, 1939). The most frequently used and intuitive component is the vertical gravity gradient, G_{zz} , which represents the rate of change of vertical gravity (g_z) with height (z). It can be deduced by differencing the value of gravity at two points separated by a small vertical distance, 1 , and dividing by this distance.

$$G_{zz} = \frac{\partial g_z}{\partial z} \approx \frac{g_z \left(z + \frac{1}{2} \right) - g_z \left(z - \frac{1}{2} \right)}{1}$$

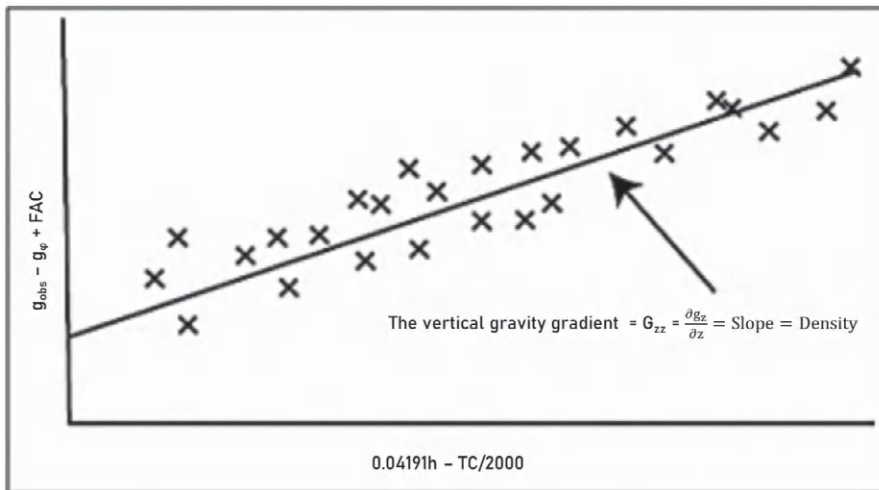


FIGURE 3.16 Rearranging the Bouguer equation.

The two gravity measurements are provided by accelerometers, which are matched and aligned to a high level of accuracy. *Density imaging* is a method that uses the inversion of the gravity and gravity gradient spectra in the wave number domain to create accurate 3-D reconstructions of subsurface density distributions. This approach offers computational efficiency and rapid calculations. Gravity and gravity gradient data are vital since they are pivotal tools that reveal the Earth's density structure and hidden mineral resources (He et al., 2024).

3.10 INTERPRETATION OF GRAVITY ANOMALIES

Data processing forms another important phase of the geophysical project. As mentioned earlier, we always attempt to determine information about the Earth's subsurface using gravity data, for which the grid of gravity values or the gravity profiles to determine the lateral location of any gravity variations are very important, requiring detailed planning in accordance with achieving the objectives of the survey. A detailed analysis is needed in order to quantify the nature (depth, geometry, density) of the subsurface feature causing the gravity variations. We have to separate the anomaly of interest (residual) from the remaining background anomaly (regional) to identify the subsurface features. Then, the residual gravity anomaly is modelled to determine the depth, density, and geometry of the anomaly's source.

3.10.1 REGIONAL AND RESIDUAL GRAVITY ANOMALIES

Over the years, several techniques have been suggested that can be used to accomplish the regional–residual anomaly separation (Telford et al., 1990). But depending on the problem—as, for example, for engineering and environmental gravity studies—the most common techniques are manual and polynomial surface fitting (Hinze, 1990). This is due to the small scale of the gravity survey, and the regional gravity field over such an area usually has little lateral changes. The simplest methods are manual techniques, such as graphical smoothing, where a simple smooth regional anomaly is subtracted from the observed gravity anomaly to obtain a residual anomaly (Figure 3.17). An advantage of the manual techniques is that the interpreter may have information on the lateral location of the source bodies, and this information can be used to select a 'correct' regional anomaly.

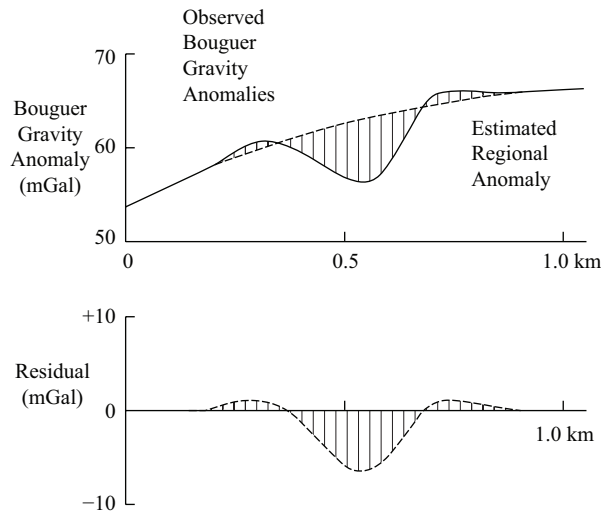


FIGURE 3.17 Example of a regional–residual gravity anomaly separation using graphical smoothing.

It is observed that the upward continuation filter is used to calculate the potential field at an elevation higher than at which the field is actually measured and is applied to both gravity and magnetic data. This filter is intended to ascertain the responses of the geologic features by progressively eliminating the effect of local features. Gravity data are observed at a level of observation at 2,000 m (Naskar and Lashkari, 2019) in the Tonk and Dausa districts, Rajasthan. A sharp gradient has been observed in several directions of the Bouguer anomaly contours, which indicates criss-cross faults that may have been caused by intensive tectonic activity, resulting in deep weathering and fracturing in the underlying basement. The gravity features, as evident from the contour pattern, are dominated by the crustal structures (Naskar and Lashkari, 2019).

It is observed that regional–residual anomaly separation techniques involve mathematical operations using a computer. One problem with the mathematical techniques is that they do not accurately represent the ‘true’ residual gravity anomaly due to a specific body. Thus, they should not be used for quantitative interpretation of the subsurface but only for qualitative interpretation (Ulrich, 1968). The most common mathematical techniques are surface fitting and weighted averaging. *Surface fitting* involves a least squares fitting of a 2-D polynomial (Beltrao et al., 1991) or 2-D Fourier series (James, 1966) of different orders to the original gridded Bouguer gravity data to represent a regional gravity anomaly map. The higher the surface order, the greater the fit to the original data; however, high-order surfaces are usually not desired, as they will contain part of the anomaly that is desired. Hinze (1990) shows a third-order polynomial surface that was removed from the original Bouguer gravity data to produce a third-order residual anomaly map over a land-fill (Hinze, 1990).

Naskar et al. (2020) interpreted a high gravity anomaly arising from Damarcherla to Gurazala in parts of the Khammam and Nalgonda districts, Telangana, that is well reflected as a high residual gravity anomaly on the residual gravity map. This elongated residual gravity high trending in nearly N–S direction and flanked by residual gravity lows on either side is inferred as basement high below overlying Kurnool sediments. Gneissic basement and overlying Banganapalle quartzite are the main pay horizons for unconformity-related uranium mineralization towards the northern and north-eastern margin of the Srisailem and Palnadu sub-basins (Naskar et al., 2020).

3.10.2 DATA ENHANCEMENT

It is very important to adopt data enhancement techniques, which are used to increase the perceptibility of the gravity anomalies, that might be related to bodies of interest. This is important in engineering and environmental gravity work, as most of the anomalies have small amplitudes and are easily obscured by the regional gravity field. The most important techniques are derivative methods. The most commonly used derivatives are the first (gradient) (Fajklewicz, 1976; Butler, 1984a, 1984b) and second (curvature) (Elkins, 1951), which are analytically calculated from a Bouguer gravity anomaly grid. It is found that the first and second derivative methods enhance near-surface anomalies at the expense of deeper anomalies and are good at locating the edges of a body. Traditionally, the second vertical derivative has been the most commonly used derivative, as the amplitude and width of a second vertical derivative are higher and narrower than the first vertical gradient and, thus, supposedly easier to interpret. It is observed that the second vertical derivative is more susceptible to data noise and errors and topographic irregularities and should only be used for large-scale interpretations. Given the problems with second vertical derivatives, numerous authors (e.g. Butler, 1984a, 1984b) developed methods of determining the vertical and horizontal gradients for shallow gravity applications. Several sets of case studies by Butler (1984b) and Mickus (1980) showed that the horizontal gravity gradients do not contain topographic effects and located shallow objects better than the vertical gravity gradients. Figure 3.18 shows observed gravity, horizontal gradient, and second vertical derivative profiles over a cavern and limestone pinnacle. Figure 3.19 shows the illustrations of the vertical gravity g , the first vertical derivative (FVD, dashed curve), the first horizontal derivative (FHD, dotted curve), and the second vertical derivative (SVD, dot-dashed curve) gravity anomaly profiles over an infinitely long 2-D horizontal cylinder striking perpendicular to the profile using normalized vertical and horizontal scales (Elkins, 1951). Figure 3.20 illustrates the vertical gravity g , the first vertical derivative (VTD), the first horizontal derivative, and the second vertical derivative (SVD) gravity anomaly profiles over an infinitely long vertical contact between geological units of differing densities, which also represent the edge of a broad anomaly source. Line styles as in Figure 3.19.

Naskar et al. (2020) showed that the second vertical derivative map (Naskar et al., 2020) clearly depicts the trend of signals forming some linear tracts with N–S alignment near the Sinapali-Larka-Lanji section in the western part of the study area (parts of the Nuapada,

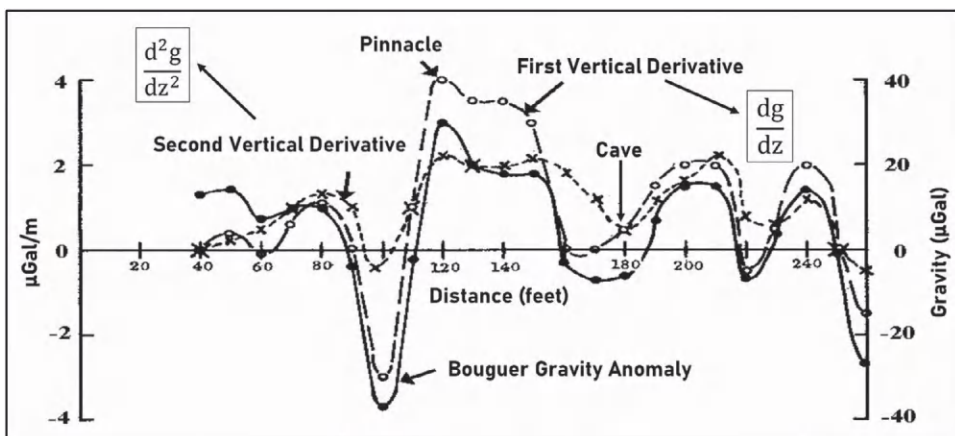


FIGURE 3.18 Bouguer gravity (solid circles), horizontal gradient (nonsolid circles), and second derivative (x's) measurements over a cave and a limestone pinnacle.

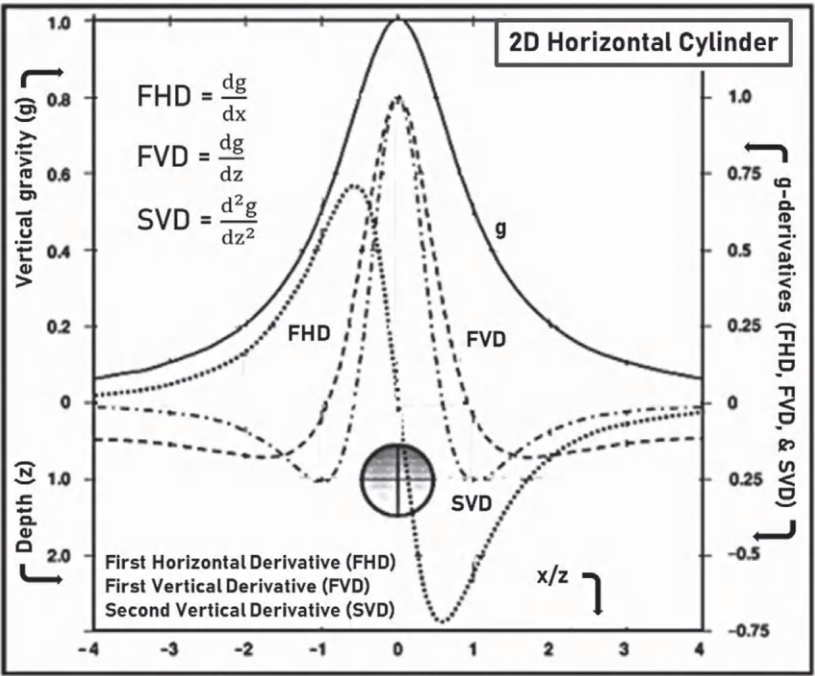


FIGURE 3.19 Signal enhancement in case of a horizontal cylinder.

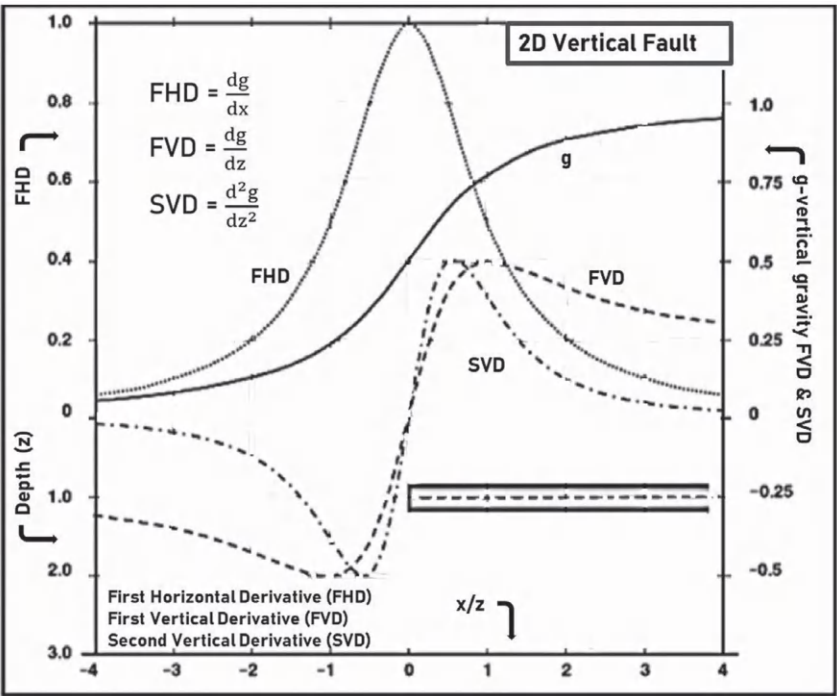


FIGURE 3.20 Signal enhancement in case of a vertical contact.

Bolangir, and Kalahandi districts, Odisha), indicating some minor contacts/faults associated with the rocks of the contact zone of Bastar Craton and EGMB. Few E–W trending linear positive anomalies, related to high-density materials, have shown enhancement near the Bhusagura, Tukla, and Gambhari areas. It resulted into an enhanced anomaly map or residual map, which has amplified anomalous signals associated with shallow features while de-emphasizing regional features.

Euler 3-D deconvolution adds an extra dimension to the interpretation of the potential field data. Mathematical testing of gravity data inferred that the Euler deconvolution method is traced back to Hood (1965) based on Euler’s homogeneity equation and derived the structural index that can be defined as a measure of the rate of change with a distance of a field. Subsequently, the study of Thompson (1982) implemented the method by applying Euler deconvolution to synthetic and real magnetic data along with profiles, and he applied this technique to magnetic data and mapped relative depth to basement, later extended by Reid et al. (1990). Since then, it has been adapted and improved upon by Keating (1998), Mushayandebvu et al. (2004), and many others. The popularity of Euler deconvolution is largely due to its great simplicity of implementation and its usability for making a quick initial interpretation. This shows that it estimates a set of (x, y, z) points that, ideally, fall inside the source of the anomaly. Euler deconvolution requires the x-, y-, and z-derivatives of the data.

According to Reid et al. (1990), the 3-D form of Euler’s equation can be defined as

$$(x - x_0) \frac{\partial T}{\partial x} + (y - y_0) \frac{\partial T}{\partial y} + (z - z_0) \frac{\partial T}{\partial z} = N(B - T)$$

where (x₀, y₀, z₀) is the position of a magnetic source whose total field *T* is detected at (x, y, z). The total field has a regional value of *B*. The degree of homogeneity *N* may be interpreted as a structural index (SI). After Reid and Thurston (2014), the values of the structural index (SI) can be as follows (Table 3.6).

Figure 3.21 shows the application of 3-D Euler deconvolution to the gravity data of the Nuapada, Bolangir, and Kalahandi districts, Odisha, India. The best solutions for the Bouguer gravity anomaly have been obtained for a grid cell size of 1,000 m, structural index (SI) of 0, window size of 10 × 10 km, and depth tolerance of 25%. The solutions have been superimposed on the combined geological and gravity contour map of the area (Naskar et al., 2020). A major number of solutions have been obtained for signal source depths ranging from 2,000 to 3,000 m and 3,000 to 4,000 m, showing depth persistence of the contacts. A well-defined linear distribution of plots of solutions, with N–S alignment in the western part of the area, possibly demarcates the deeper part of contact of Bastar Craton and EGMB. At this depth level, the contact shows prominent density contrast with the surrounding medium. The depth of the deeper part of the contact varies mostly between 1,000 and 4,000 m, with a plunge towards the north. Khondalite–gneissic contact is deeper than 4,000 m near Lathor, and 2,000 m to the south-east of Kantabanji.

TABLE 3.6
Structural Index

Source Type	SI (Mag)	SI (Grav)
Point, sphere	3	2
Line, cylinder, thin bed fault	2	1
Thin sheet edge, thin sill, thin dyke	1	0

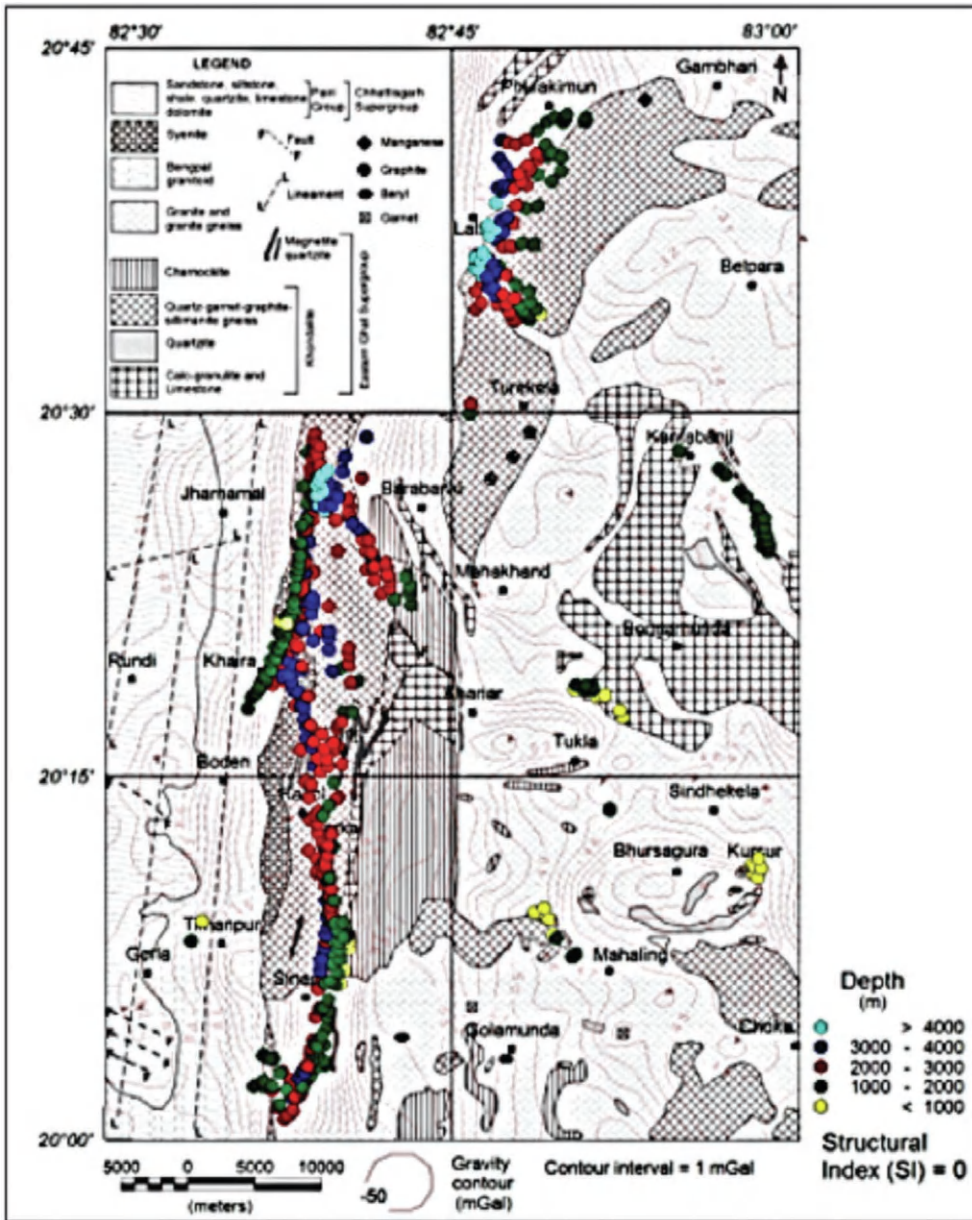


FIGURE 3.21 Euler depth solutions from Bouguer gravity anomaly, parts of the Nuapada, Bolangir, and Kalahandi districts, Odisha.

3.10.3 INTERPRETATIONAL TECHNIQUES IN THE INVERSE PROBLEM

Two ways of solving the inverse problem: direct and indirect interpretation (Figure 3.22; EarthDoc).

3.10.3.1 Direct Interpretation

Assumption: a 3-D anomaly is caused by a point mass (a 2-D anomaly is caused by a line mass) at depth = z , where x_1^2 gives 'z' (Figure 3.23; Telford et al., 1990).

Following, the direct data interpretation table (Table 3.7) is given for easy comprehension.

3.10.3.2 Indirect Interpretation

The following steps are required for the indirect interpretation of gravity anomalies (Figures 3.24a and b; Radogna et al., 2004):

- Construction of a reasonable model
- Computation of its gravity anomaly
- Comparison of computed with observed anomaly
- Alteration of the model to improve correspondence of observed and calculated anomalies and return to second step

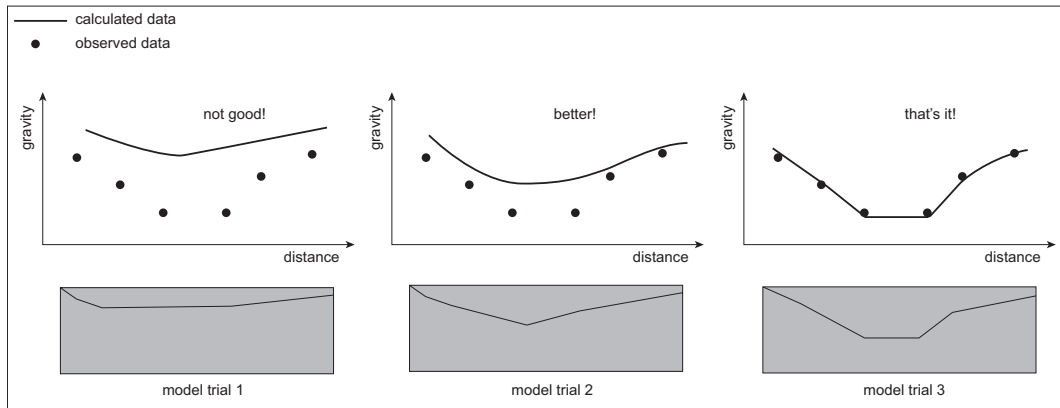


FIGURE 3.24A Steps for indirect interpretation.

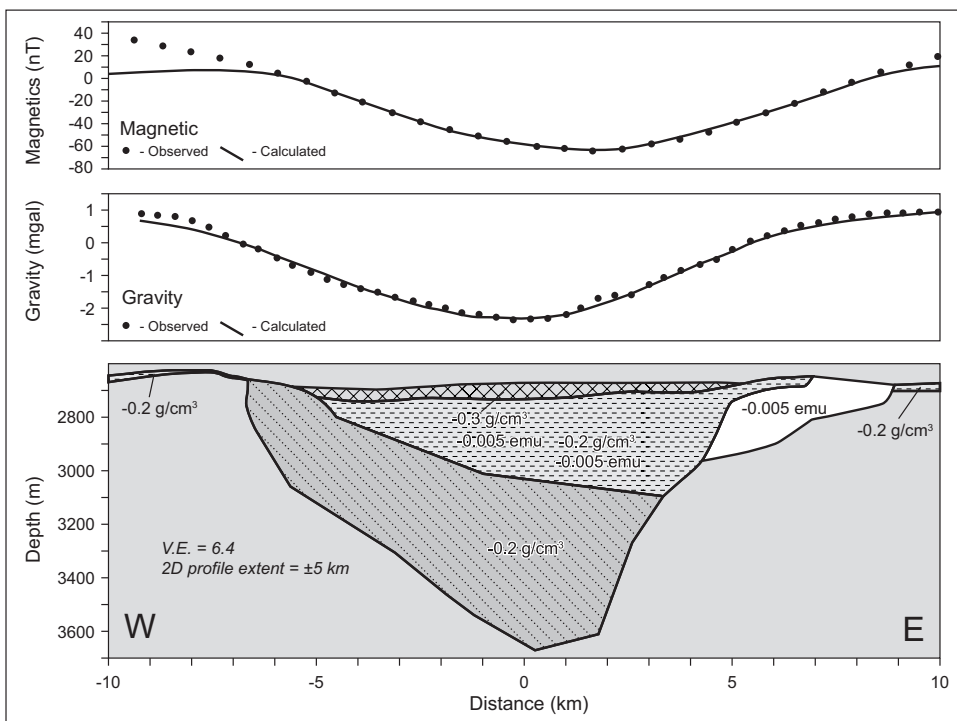


FIGURE 3.24B Final step for indirect interpretation.

3.10.3.3 Non-Unicity of the Solution

It is important to mention that the ‘non-uniqueness’ of potential field studies is closely related to the more general topic of scientific uncertainty in the Earth sciences and beyond. Several studies demonstrated that all results in the Earth sciences are subject to significant uncertainty because problems are generally addressed with incomplete and imprecise data. We therefore need to combine results from multiple disciplines into the integrated solutions for addressing complex global issues requiring special attention to the appreciation and communication of uncertainty in geologic interpretation.

A basic illustration of the non-uniqueness of unconstrained potential field interpretation is the construction of a set of discrete bodies, each of which yields the same calculated anomaly (Figure 3.25; Radogna et al., 2004).

3.10.4 MODELLING GRAVITY ANOMALIES

We are always concerned with the removal of regional effects to get the residual gravity anomaly, which must be interpreted in terms of an anomalous density distribution. Modern analyses are based on iterative modelling using high-speed computers. Previous studies showed that earlier methods of interpretation utilized the comparison of the observed gravity anomalies with the computed anomalies related to the different geometric shapes. However, the success of this simple approach is seen mainly due to the insensitivity of the shape of a gravity anomaly to minor variations in the anomalous density distribution. Moreover, several other fundamental problems of interpreting gravity anomalies can be learned from the computed effects of geometric models. Thus, justifiable is the observation that the interpretation of gravity anomalies is not unique since different density distributions can give the same anomaly and so may confuse the interpretation to take decision which gravity anomaly corresponds to which of the subsurface bodies.

3.10.4.1 Gravity Anomaly of a Point Mass or Sphere: Model for a Diapir

Interestingly, diapiric structures contain materials of a different density into the host rock. A low-density salt dome ($\rho = 2,150 \text{ kg m}^{-3}$) intruding upon higher-density carbonate rocks ($\rho_0 = 2,500 \text{ kg m}^{-3}$) has a density contrast $\delta\rho = -350 \text{ kg m}^{-3}$ and causes a negative gravity anomaly. A volcanic

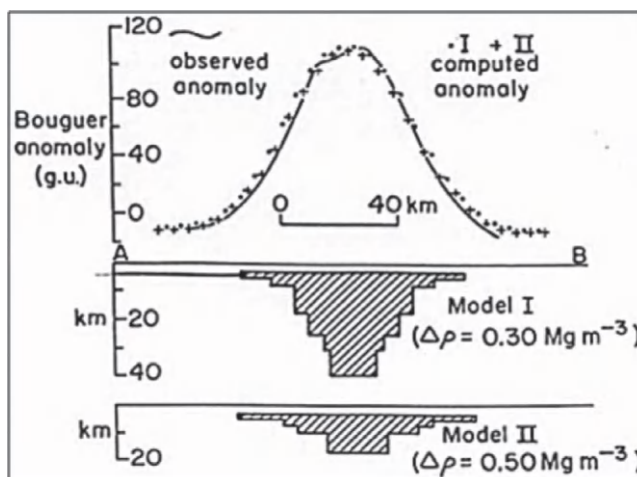


FIGURE 3.25 Representation of 2-D gravity modelling.

plug ($\rho = 2,800 \text{ kg m}^{-3}$) intruding upon a granite body ($\delta\rho = 2,600 \text{ kg m}^{-3}$) has a density contrast $\delta\rho = +200 \text{ kg m}^{-3}$, which causes a positive gravity anomaly. The contour lines on a map of the anomaly are centred on the diapir, so all profiles across the centre of the structure are equivalent. The anomalous body can be modelled equally by a vertical cylinder or by a sphere, which we will evaluate here because of the simplicity of the model.

Assume a sphere of radius R and density contrast $\delta\rho$ with its centre at depth z below the surface (Figure 3.26). The attraction δg of the sphere is as though the anomalous mass M of the sphere were concentrated at its centre. If we measure the horizontal position from a point above its centre, at distance x , the vertical component δg_z is given by

$$\delta g_z = \delta g \sin\theta = \frac{GM}{r^2} \frac{z}{r} \quad (3.27)$$

where $M = \frac{4}{3}\pi R^3 \delta\rho$ and $r^2 = z^2 + x^2$.

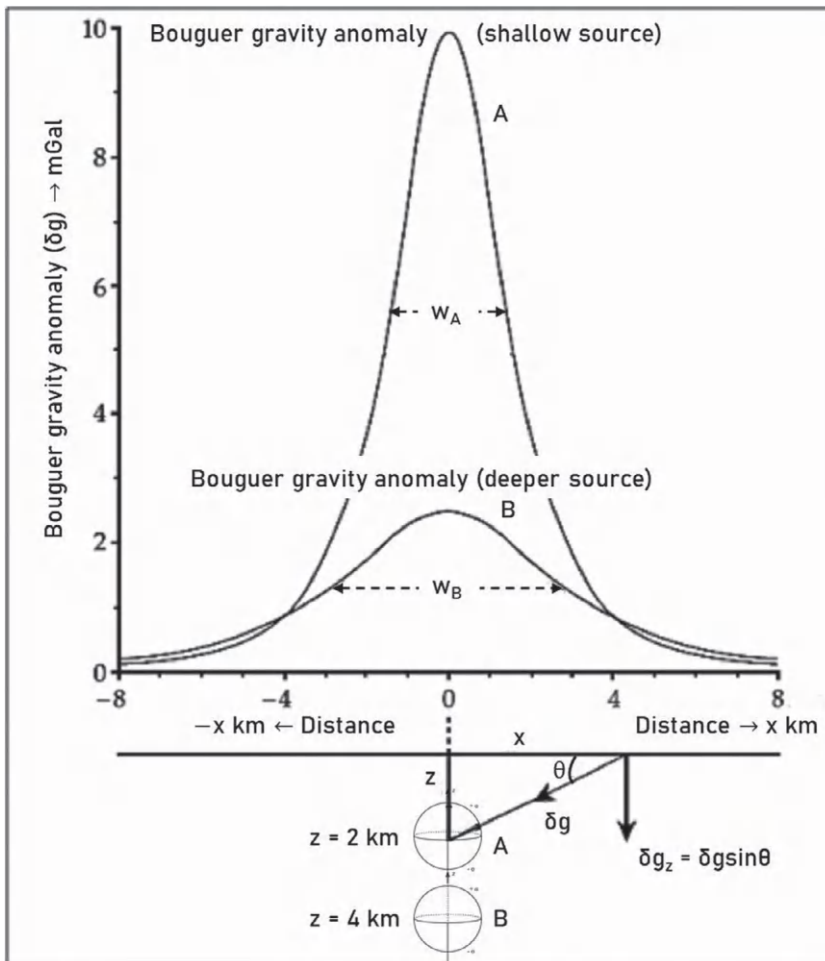


FIGURE 3.26 Gravity anomalies for buried spheres with the same radius R and density contrast $\delta\rho$, but with their centres at different depths z below the surface. The anomaly of the deeper sphere B is flatter and broader than the anomaly of the shallower sphere.

Substituting these expressions into eq. 3.27 and rearranging the terms give

$$\delta g_z = \frac{4}{3} \pi G \delta \rho R^3 \frac{z}{(z^2 + x^2)^{\frac{3}{2}}} = \frac{4}{3} \pi G \left(\frac{\delta \rho R^3}{z^2} \right) \frac{1}{\left[1 + \left(\frac{x}{z} \right)^2 \right]^{\frac{3}{2}}} \quad (3.28)$$

The terms in the first pair of parentheses depend on the size, depth, and density contrast of the anomalous sphere. They determine the maximum amplitude of the anomaly, δg_0 , which is reached over the centre of the sphere at $x = 0$. The peak value is given by

$$\delta g_0 = \frac{4}{3} \pi G \left(\frac{\delta \rho R^3}{z^2} \right) \quad (3.29)$$

This equation shows how the depth to the centre of the sphere affects the peak amplitude of the anomaly; the greater the depth to the centre, the smaller the amplitude (Figure 3.26; Saibi, 2018). For a given depth, the same peak anomaly can be produced by numerous combinations of $\delta \rho$ and R ; a large sphere with a low-density contrast can give an identical anomaly to a small sphere with a high-density contrast (Talwani et al., 1959). The gravity data alone do not allow us to resolve this ambiguity (Section 3.9.3.3 earlier).

The terms in the second pair of parentheses in eq. 3.27 describe how the amplitude of the anomaly varies with distance along the profile. The anomaly is symmetrical with respect to x , reaches a maximum value δg_0 over the centre of the sphere ($x = 0$), and decreases to 0 at great distances ($x = \infty$). Note that the larger the depth z , the more slowly the amplitude decreases laterally with increasing x . A deep source produces a smaller but broader anomaly than the same source at shallower depth. It is observed that the width w of the anomaly where the amplitude has one-half its maximum value is called the *half-height width*. The depth z to the centre of the sphere is deduced from this anomaly width from the relationship $z = 0.652 w$.

3.10.4.2 Horizontal Line Element

Many geologically interesting structures extend to great distances in one direction but have the same cross-sectional shape along the strike of the structure. If the length along the strike were infinite, the 2-D variation of density in the area of the cross section would suffice to model the structure. But the lateral extent is never infinite, so this is not really valid. As a general rule, if the length of the structure normal to the profile is more than 20 times its width or depth, it can be treated as 2-D. The limited lateral extent of the structure must be considered in computing its anomaly because of the end effect. It is interesting to note that an elongated body that requires end corrections is sometimes referred to as a 2.5D structure. For example, the mass distribution under elongated bodies like anticlines, synclines, and faults should be modelled as 2.5D structures. Here, we will handle the simpler 2-D models of these structures.

Let an infinitely long linear mass distribution with mass m per unit length extend horizontally along the y -axis at depth z (Figure 3.27). The contribution d (δg_z) to the vertical gravity anomaly δg_z at a point on the x -axis due to a small element of length dy is

$$d(\delta g_z) = \frac{G m dy}{r^2} \sin \theta = \frac{G m dy}{r^2} \frac{z}{r} \quad (3.30)$$

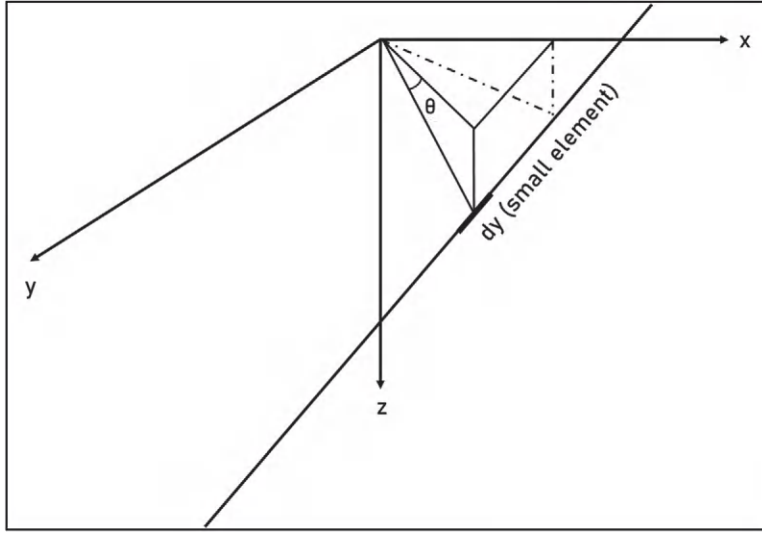


FIGURE 3.27 Geometry for calculation of the gravity anomaly of an infinitely long linear mass distribution with mass m per unit length extending horizontally along the y -axis at depth z .

Since the line element extends to infinity along the positive and negative y -axis, its vertical gravity anomaly is found by integration:

$$\delta g_z = Gmz \int_{-\infty}^{+\infty} \frac{dy}{r^3} = Gmz \int_{-\infty}^{+\infty} \frac{dy}{(u^2 + y^2)^{\frac{3}{2}}} \quad (3.31)$$

where $u^2 = x^2 + z^2$. The integration is simplified by changing variables so that $y = u \tan \varphi$; then $dy = u \sec^2 \varphi \, d\varphi$ and $(u^2 + y^2)^{\frac{3}{2}} = u^3 \sec^3 \varphi$. This gives

$$\delta g_z = \frac{Gmz}{u^2} \int_{-\frac{\pi}{2}}^{+\frac{\pi}{2}} \cos \varphi \, d\varphi \quad (3.32)$$

which, after evaluation of the integral, gives

$$\delta g_z = \frac{2Gmz}{z^2 + x^2} \quad (3.33)$$

The preceding expression can be written as the derivative of a potential function Ψ :

$$\delta g_z = Gm \frac{2z}{u^2} = - \frac{\partial \Psi}{\partial z} \quad (3.34)$$

$$\Psi = Gm \log_e \left(\frac{1}{u} \right) = Gm \log_e \frac{1}{\sqrt{x^2 + z^2}} \quad (3.35)$$

Ψ is called the logarithmic potential. Eqs. 3.33 and 3.35 are useful results for deriving formulas for the gravity anomaly of linear structures like an anticline (or syncline) or a fault.

3.10.4.3 Gravity Anomaly of a Horizontal Cylinder: Model for Anticline or Syncline

We observed that the gravity anomaly of an *anticline* can be modelled by assuming that the upward folding of strata brings rocks with higher density nearer to the surface (Figure 3.28, left; Saibi, 2018), thereby causing a positive density contrast, whilst a *syncline* is modelled by assuming that its core is filled with strata of lower density that cause a negative density contrast. In each case, the geometric model of the structure is an infinite horizontal cylinder (Figure 3.28, right).

It is evident that a horizontal cylinder is composed of numerous line elements parallel to its axis. The cross-sectional area of an element (Figure 3.29; Lowrie, 2007) gives a mass anomaly per unit length, $m = \delta\rho r d\theta dr$. The contribution $d\Psi$ of a line element to the potential at the surface is

$$d\Psi = 2G \delta\rho \log_e \left(\frac{1}{u} \right) r dr d\theta \quad (3.36)$$

It shows that integrating over the cross section of the cylinder gives its potential Ψ ; the vertical gravity anomaly of the cylinder is then found by differentiating Ψ with respect to z . Noting that $\frac{du}{dz} = \frac{z}{u}$, we get

$$\delta g_z = - \frac{\partial \Psi}{\partial z} = - \frac{z}{u} \frac{\partial \Psi}{\partial u} = - \frac{2G\delta\rho z}{u} \int_0^{2\pi} \int_0^R \frac{\partial}{\partial u} \log_e \left(\frac{1}{u} \right) r dr d\theta \quad (3.37)$$

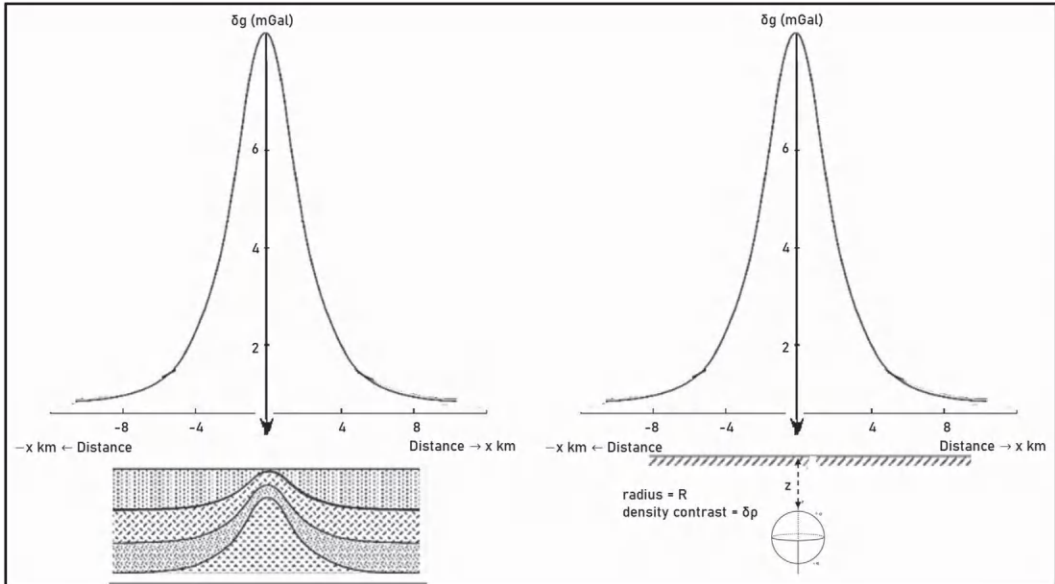


FIGURE 3.28 Calculation of the gravity anomaly of an anticline: (left) structural cross section, and (right) geometric model by an infinite horizontal cylinder.

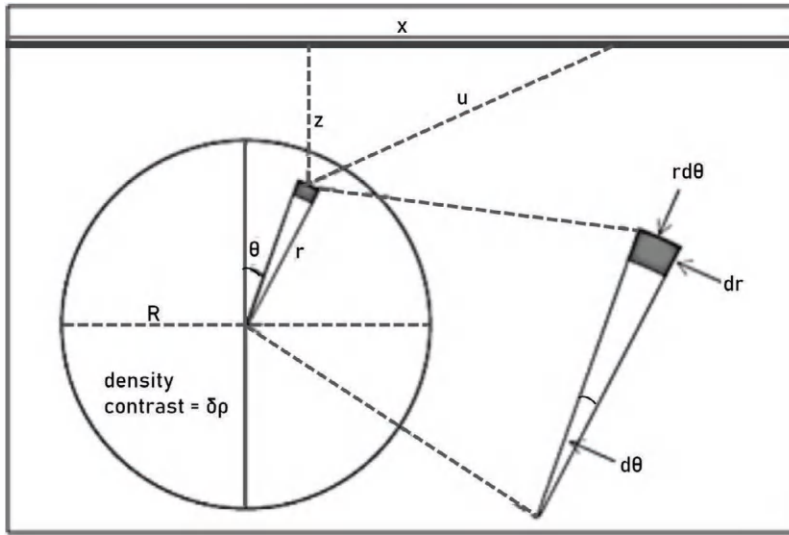


FIGURE 3.29 Cross-sectional geometry for calculating the gravity anomaly of a buried horizontal cylinder made up of line elements parallel to its axis.

After first carrying out the differentiation within the integral, this simplifies to

$$\delta g_z = \frac{2G\delta\rho z}{u^2} \int_0^{2\pi} \int_0^R r dr d\theta = \frac{2\pi G R^2 \delta\rho z}{x^2 + z^2} \quad (3.38)$$

Comparing eq. 3.38 and eq. 3.33, it is evident that the gravity anomaly of the cylinder is the same as that of a linear mass element concentrated along its axis, with mass $m = \pi R^2 \delta\rho$ per unit length along the strike of the structure. The anomaly can be written as

$$\delta g_z = 2\pi G \left(\frac{\delta\rho R^2}{z} \right) \left[\frac{1}{1 + \left(\frac{x}{z} \right)^2} \right] \quad (3.39)$$

The shape of the anomaly on a profile normal to the structure (Figure 3.28, right) resembles that of a sphere (Figure 3.26). The central peak value δg_0 is given by

$$\delta g_0 = 2\pi G \left(\frac{\delta\rho R^2}{z} \right) \quad (3.40)$$

The anomaly of a horizontal cylinder decreases laterally less rapidly than that of a sphere due to the long extent of the cylinder normal to the profile. It is observed that 'half-height width' w of the anomaly is again dependent on the depth z to the axis of the cylinder; in this case, the depth is given by $z = 0.5 w$.

3.10.4.4 Horizontal Thin Sheet

We next compute the anomaly of a thin horizontal ribbon of infinite length normal to the plane of the profile. This is done by fictively replacing the ribbon with numerous infinitely long line

elements laid side by side. Let the depth of the sheet be z , its thickness t , and its density contrast $\delta\rho$ (Figure 3.30a); the mass per unit length in the y -direction of a line element of width dx is $(\delta\rho t \, dx)$. Substituting in $\delta g_z = \frac{2Gmz}{z^2 + x^2}$, we get the vertical gravity anomaly of a horizontal line element; the anomaly of the thin ribbon is then computed by integrating between the limits x_1 and x_2 (Figure 3.30b):

$$\delta g_z = 2G \delta\rho t z \int_{x_1}^{x_2} \frac{dx}{x^2 + z^2} = 2G \delta\rho t \left[\tan^{-1} \left(\frac{x_2}{z} \right) - \tan^{-1} \left(\frac{x_1}{z} \right) \right] \quad (3.41)$$

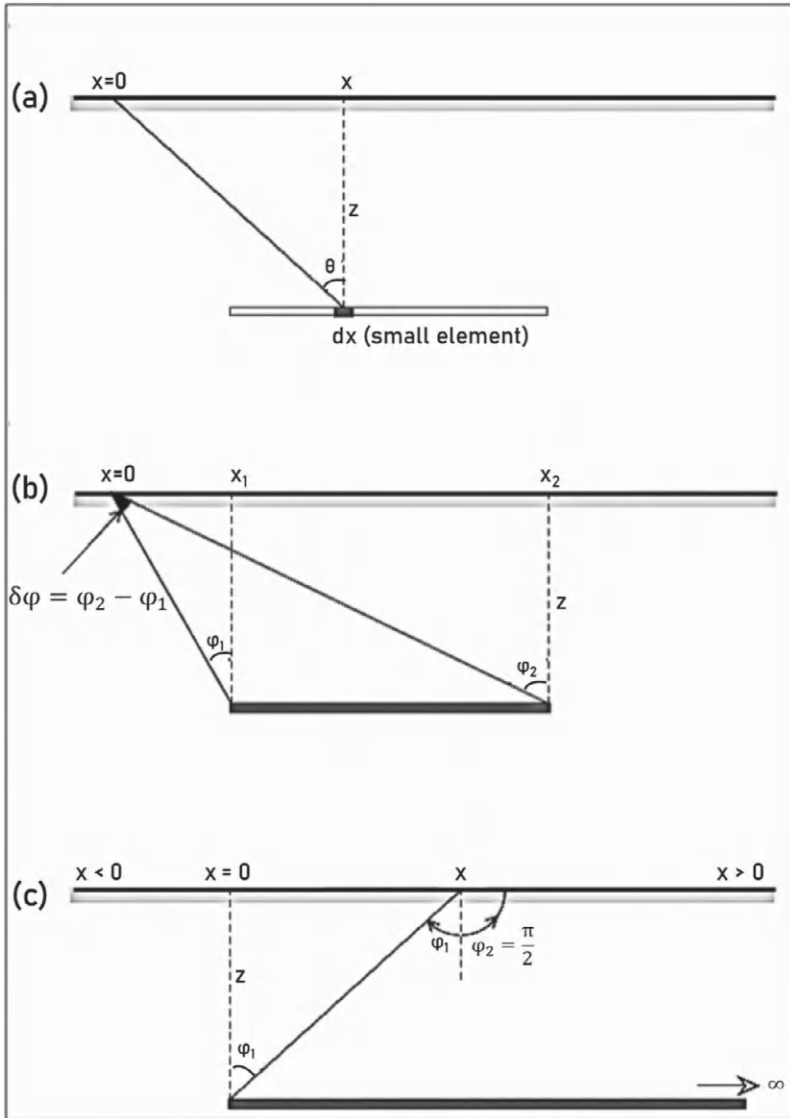


FIGURE 3.30 Geometry for computation of the gravity anomaly across a horizontal thin sheet: (a) subdivision of the ribbon into line elements of width dx , (b) thin ribbon between the horizontal limits x_1 and x_2 , and (c) semi-infinite horizontal thin sheet.

Writing $\tan^{-1}\left(\frac{x_1}{z}\right) = \varphi_1$ and $\tan^{-1}\left(\frac{x_2}{z}\right) = \varphi_2$, as in Figure 3.30b, the equation becomes

$$\delta g_z = 2G \delta g_z t (\varphi_2 - \varphi_1) \quad (3.42)$$

It should be remembered that the gravity anomaly of the horizontal ribbon is proportional to the angle it subtends at the point of measurement.

The anomaly of a semi-infinite horizontal sheet is a limiting case of this result. For easier reference, the origin of x is moved to the edge of the sheet so that distances to the left are negative and those to the right are positive (Figure 3.30c).

This makes $\varphi_1 = -\tan^{-1}\left(\frac{x}{z}\right)$. The remote end of the sheet is at infinity, and $\varphi_2 = \frac{\pi}{2}$. The gravity anomaly is then

$$\delta g_z = 2G \delta \rho t \left[\frac{\pi}{2} + \tan^{-1}\left(\frac{x}{z}\right) \right] \quad (3.43)$$

It is a good example that the infinite horizontal sheet extends to infinity in the positive and negative x and y directions. With $\varphi_2 = \frac{\pi}{2}$ and $\varphi_1 = -\frac{\pi}{2}$, the anomaly is

$$\delta g_z = 2\pi G \delta \rho t \quad (3.44)$$

The preceding equation is the same as the expression for the Bouguer plate correction.

3.10.4.5 Horizontal Slab: Model for a Vertical Fault

The gravity anomaly across a vertical fault increases progressively to a maximum value over the uplifted side (Figure 3.31a). This is interpreted as due to the upward displacement of denser material, which causes a horizontal density contrast across a vertical step of height h (Figure 3.31b). The faulted block can be modelled as a semi-infinite horizontal slab of height h and density contrast $\delta \rho$, with its midpoint at depth z_0 (Figure 3.31c).

Let us suppose that the slab is divided into thin, semi-infinite horizontal sheets of thickness dz at depth z . The gravity anomaly of a given sheet is given by $\delta g_z = 2G \delta \rho t \left[\frac{\pi}{2} + \tan^{-1}\left(\frac{x}{z}\right) \right]$, with dz for the thickness (t). The anomaly of the semi-infinite slab is found by integrating with respect to z over the thickness of the slab; the limits of integration are $z - \left(\frac{h}{2}\right)$ and $z + \left(\frac{h}{2}\right)$. After slightly rearranging terms, this gives

$$\delta g_z = 2G \delta \rho h \left[\frac{\pi}{2} + \frac{1}{h} \int_{z_0 - \frac{h}{2}}^{z_0 + \frac{h}{2}} \tan^{-1}\left(\frac{x}{z}\right) dz \right] \quad (3.45)$$

It is observed that the second expression in the brackets is the mean value of the angle $\tan^{-1}\left(\frac{x}{z}\right)$ averaged over the height of the fault step. This can be replaced to a good approximation by the value at the midpoint of the step, at depth z_0 . This gives

$$\delta g_z = 2G \delta \rho h \left[\frac{\pi}{2} + \tan^{-1}\left(\frac{x}{z_0}\right) \right] \quad (3.46)$$

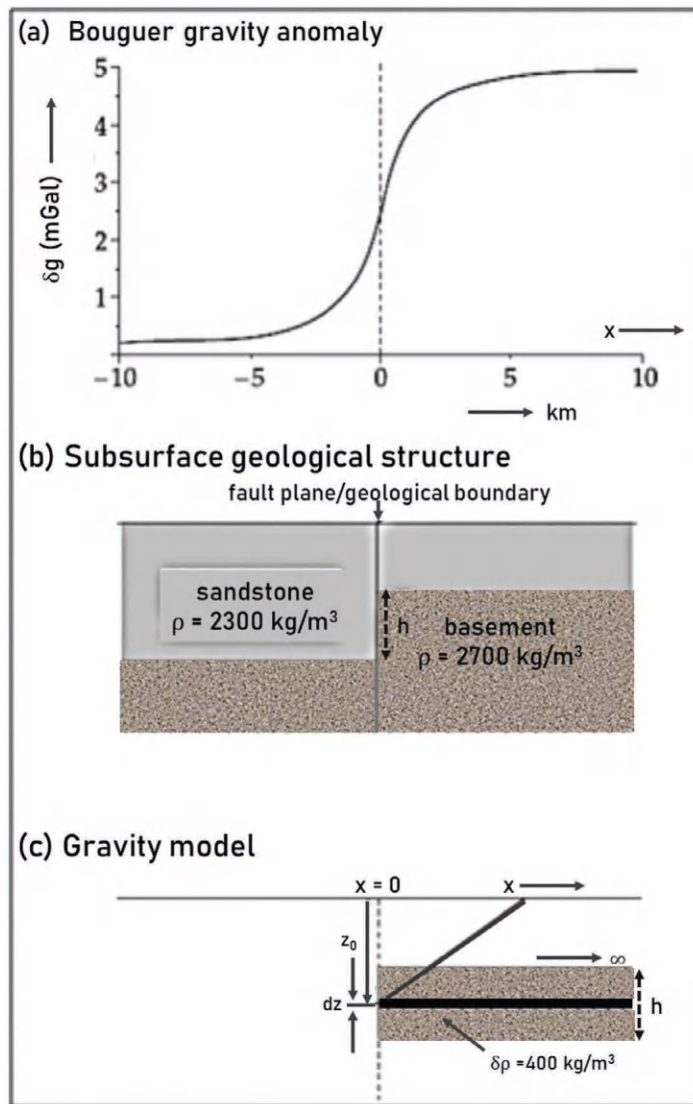


FIGURE 3.31 (a) The gravity anomaly across a vertical fault, (b) the structure of a fault with vertical displacement h , and (c) the model of the anomalous body as a semi-infinite horizontal slab of height h .

A comparison of this expression with eq. (3.43) shows that the anomaly of the vertical fault (or a semi-infinite thick horizontal slab) is the same as if the anomalous slab were replaced by a thin sheet of thickness h at the midpoint of the vertical step. Eq. (3.46) is called the ‘thin sheet approximation’. It is accurate to about 2%, provided that $z_0 > 2h$.

3.11 APPLICATIONS OF GRAVITY METHOD AND CASE STUDY

As mentioned in the beginning, the gravity method of geophysical exploration is based on the measurement of variations in the gravity field caused by horizontal variations of density within the subsurface. A series of study on gravimetry demonstrated that it is an important technique for many problems that involve subsurface mapping, and it is the principal method in a number of specific

types of geological studies. The method, which has its roots in geodetic studies from the 17th to the 20th centuries, has developed rapidly over the past century as a result of significant technological advances: primarily high-accuracy gravity measuring instruments for use on land, sea, and air, and in space, as well as the increasing computational and graphics power of digital computers. It is worth mentioning that these advances are backed up by continuing improvements in data processing, interpretational schemes, practical experience, and surveying methodologies.

Usually, the gravity method has been used primarily in regional characterization of the Earth for determining the architecture of the crust, identifying potentially favourable regions for resource exploration, and developing conceptual exploration models (Hinze, 1990, 2013). This is made possible by the millions of gravity observations now available worldwide in both public and commercial datasets. Still, the accuracy now available in gravity observations and the relative ease of measurement have made the method viable for exploration objectives, such as assessing subsurface changes over time and combined interpretation with seismic reflection mapping. It has also been possible to delineate suspected subsurface voids by gravity method characterized by residual gravity lows that appear to be most vulnerable to subsidence (Biswas et al., 1999, 2005).

3.11.1 GRAVITY SURVEY ACROSS THE NARMADA–SON LINEAMENT (NSL), CENTRAL INDIA, MP

A 2-D gravity model along a profile across the Narmada–Son lineament (NSL) in Central India (Naskar et al., 2008a), with a good amount of overlap on both sides. The location of places, faults, the different tectonic subdivisions, and the interpreted crustal layers along with densities of crustal rocks are shown. It is observed that the Moho depth north of the NSL is comparatively less than what is interpreted over the concealed Vindhyan basin below the trap, which is around 37 km. It is corroborated by Kaila et al. (1985) through their deep seismic sounding (DSS) work. The Moho interpreted from DSS has also been incorporated in this model (Naskar et al., 2008b). There is a substantial Moho upwarp to the north of the Narmada River and around the Satpura horst structure south of the Narmada River. The depth to the Conrad is modelled between 12 km and 20 km. Patches of Mahakoshals (density 2.95 gm/cc) below the Vindhyan cover have been interpreted near the Indore-Choral sector, north of Ujjain, and below the Satpura horst. South of Sanwad in Borgaon and the Burhanpur area, a lithounit having a density of 2.40 gm/cc has been interpreted, which is analogous to the density of the Gondwana lithopackage. In all probability, the Gondwana exposed north of Betul may continue below the trap in this region. The thicknesses of the Deccan trap and the Vindhyan are nearly 1 km and 3 km, respectively. The Mahakoshal package in the Satpura region has a thickness between 3 km and 4 km. One deep fault which affects both Conrad and Moho has been inferred from the sharp change in the gradient of the respective boundaries to the immediate north of the Narmada River, which marks the northern boundary of the rift structure. The Moho upwarp below the Narmada River is a characteristic feature of the Narmada rift structure.

A 2-D gravity model along the profile BB' is shown in Figure 3.32. It crosses the Vindhyan metasediments and the Quaternaries near Jabalpur and passes through the trap rock down south around Lakhnadon and enters the Archaean (Naskar et al., 2008b). The Moho depth is shallowest near Jabalpur, where an upwarping in the Moho morphology is prominent, which spreads over a distance of almost 120 km from south of Damoh to Lakhnadon (Figure 3.32). Possibly, this upwarping in the Moho created the Proterozoic rift on which the Mahakoshals and Vindhyan, with a composite thickness of almost 3–4 km, have been deposited. The interpreted thickness of trap is about 1 km to 1.5 km south of Jabalpur. It is found that the Conrad is again very much variable to the south of Jabalpur due to possible crustal faults reaching up to that depth, which controlled the geological architecture of this area. This is supported by the steep gradient of the gravity profile at several places. However, the presence of Mahakoshals above the felsic granites to the south of Jabalpur is not

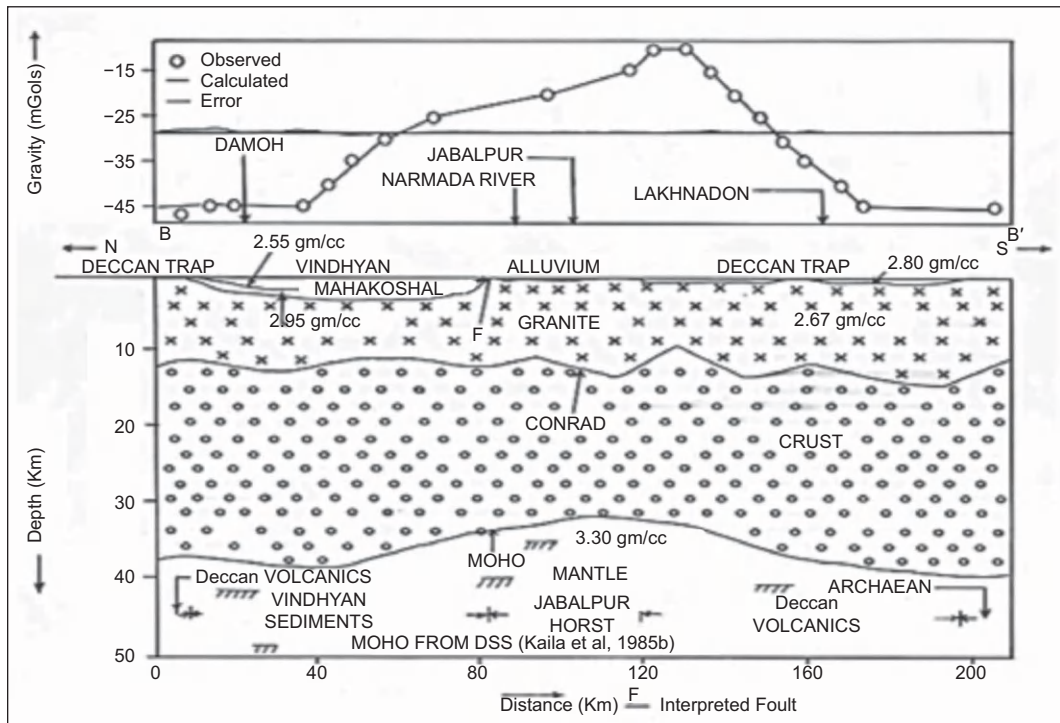


FIGURE 3.32 A 2-D gravity model along the BB' profile (Damoh–Jabalpur and east of Seoni), Central India, MP.

ruled out. In the absence of any direct evidence, this has not been shown in the 2-D model in this part. The highest Bouguer anomaly value between Jabalpur and Lakshadon may either be due to upwarping of the Conrad or due to a 3 km to 4 km package of Mahakoshals at higher stratigraphic level. The fault (F–F) to the immediate north of the Narmada River at the southern contact of the Vindhyan corroborates with the earlier MT (Gokarn et al., 2001) and magnetic data (Naskar et al., 2003).

3.11.2 GRAVITY SURVEY IN LESSER HIMALAYAS ALONG THE PURNEA–SEVOKE TRANSECT

The study of crustal structure in the sub- and lesser Himalaya regions has always attracted the attention of geoscientists because of its academic and economic importance. A classical study on gravity has been carried out in the lesser Himalaya along the Purnea–Sevoke transect to delineate the thicknesses of the lower Siwaliks/Gondwana formations above the basement, and to prepare a meaningful crustal model, the area bounded by latitudes 25°52' N to 27° N and longitudes 87°28' E to 88°28' E was surveyed using the gravity technique (Naskar, 2016). In order to understand the variations of Bouguer anomaly in terms of crustal structure, that is, mass distribution, 2-D gravity modelling was done over the Purnea–Sevoke gravity profile (Figure 3.33). A four-layer crustal model was assumed, keeping in view the local geology and the intercepts from the frequency analysis data. The gravity value was calculated after Talwani et al. (1959). Values of individual crustal layers are shown in the figure. The thickness of the composite sedimentary column has been varied from 0.5 km to 5.5 km from south-west to north-east (Figure 3.33). It conspicuously revealed the two basinal features associated with equal numbers of basement rise, which are interpreted by Naskar (2016). Several faults at km st 425, 440, 455, 480, and 520 are also interpreted from the 2-D gravity modelling.

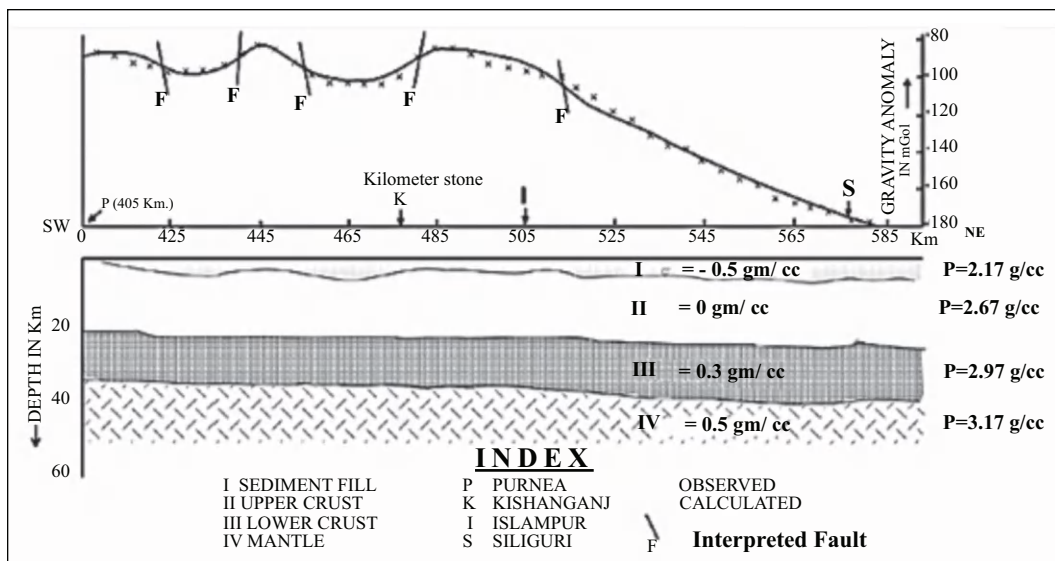


FIGURE 3.33 Bouguer anomaly profile and 2-D gravity model from Purnea to Sevoke along NH-31: $\sigma = -0.5$ g/cc represents density contrast. $P = 2.17$ g/cc indicates density of the layer.

3.11.3 THE USE OF GRAVITY METHODS IN THE INTERNAL CHARACTERIZATION OF LANDFILLS: A CASE STUDY

Landfills are a classical way of waste disposal because of their complexity and the heterogeneity that they pose, challenging problems to geophysics (Reynolds, 1997a, 1997b). Geophysics has been used to study contamination plumes originating from landfills, and the extension both on the surface and in the depth of plumes, as well as their evolution with time, has been studied (Benson et al., 1983; Greenhouse et al., 1989; Matias et al., 1994).

Gravity has already been used in landfill exploration. The effectiveness of the method to determine lateral and vertical boundaries of landfills was investigated by Roberts et al. (1990) and Silva et al. (2008). The gravity method has proven its potential in investigating whether it is possible to distinguish between different types of waste and to map important variations in bulk density of a landfill site (Galland and Stierman, 2005). Herein, a case study of the application of gravity methods of the Ilhavo (a municipality located in the centre of Portugal) landfill to the internal characterization of a sealed landfill is presented (Mantlik et al., 2009).

The gravity survey consisted of two parallel north–south profiles, *P1* and *P2*, depicted in Mantlik et al. (2009). Profile *P1*, length 1,050 m, crosses the centre of the landfill, and its aims are to investigate both the lateral and vertical boundaries of the landfill, as well as the possible blocks of different density inside the landfill. The complete Bouguer anomalies at the main profile and the control profile are depicted in Mantlik et al. (2009). It is observed that the anomaly shape at the profile *P2* represents the influence of the regional effects. In contrast, the complete Bouguer anomaly in profile *P1* represents the superposition of the regional effects and the gravity effect of the landfill deposits. If regional effects are removed, the gravity field influence of the landfill masses can be separated. The resulting residual fields obtained from both methods (standard method: linear polynomial fitting; differential method: subtraction of the regional component using *P2* readings) are depicted in Mantlik et al. (2009).

As can be seen, the difference between the residual fields computed by the two methods is negligible in the region over the surface of the waste deposits. Moreover, another interesting feature

of the use of control profile P2 is the absence of the small positive anomaly and the significant lowering of another small anomaly to the north of the landfill after subtracting the P2 readings from the P1 values. The residual field is shown in Mantlik et al. (2009). Analysis of this figure allows a qualitative interpretation of the gravity anomaly over the landfill. Hence, there are three different zones within the landfill: a short-period positive gravity anomaly up to $+50 \mu\text{Gal}$ is located in the northern part of the landfill; a second, low-magnitude positive anomaly less than $+10 \mu\text{Gal}$ in the central part; and finally, a third significant negative anomaly about $-50 \mu\text{Gal}$ in the southernmost part.

The complete interpretation of the anomaly data will be carried out with the aid of a priori information provided by 2-D resistivity data. It has been reported that the resistivity survey, together with borehole information, was used to constrain the gravity model. Combined analyses estimated the landfill thickness value to 8 m, and the lateral limits of the landfill were also estimated using the gravity profile. The residual gravity response was modelled using the GM-SYS software and is plotted in Figure 3.34. The regional density value used for the modelling was 2.517 gm/cc (Mantlik et al., 2009).

The model shown in Figure 3.34 was obtained in accordance with the resistivity data model. The high-density deposits of 2.7 gm/cc at the northern part of the landfill correspond to either the oldest layers, possibly with high compaction, or a landfill area where bulk materials had been deposited. On the other hand, the region of the landfill that shows low density of 2.35 gm/cc material could correspond either to the youngest part of the landfill or to a region where the fluid phase prevails. The gravity block could be caused by lower compaction of the younger deposits. However, inhomogeneities or hydrological effects could play an important role as well. In the middle part, the density of the deposits of 2.58 gm/cc is not significantly different from the background sand with estimated density of 2.52 gm/cc . Possible lateral change in the character of

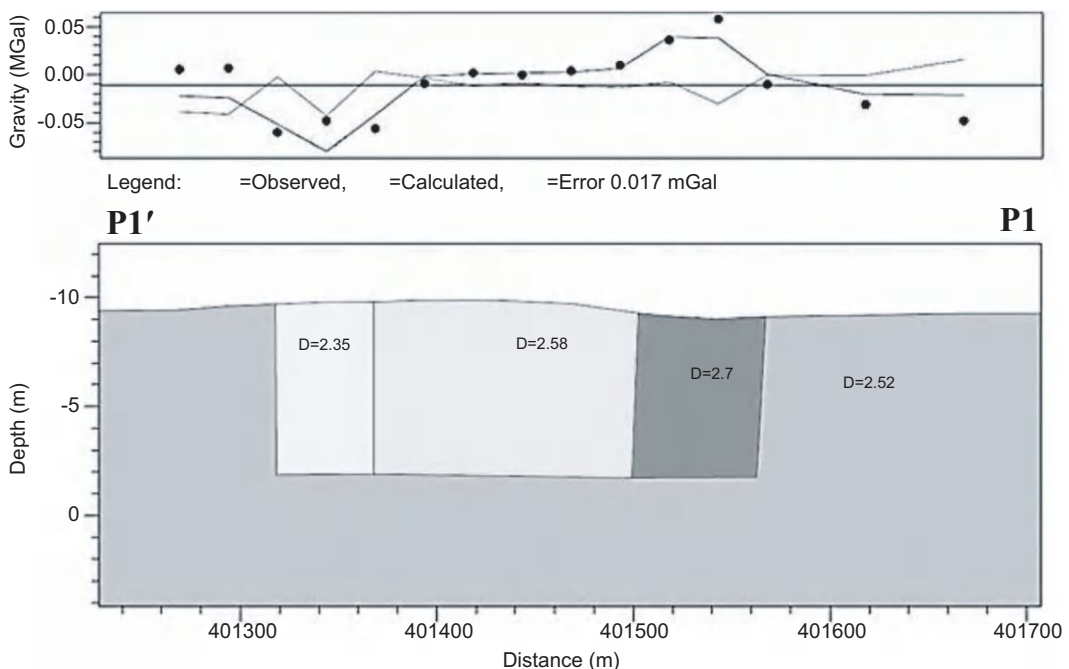


FIGURE 3.34 Gravity model of the Ilhavo landfill. The depth of the landfill deposits was determined from the resistivity data interpretation that allowed estimating the overall densities of the three zones inside the landfill.

the landfill deposits approximately 60–70 m from the northern landfill boundary in the gravity model (Figure 3.34) corresponds to the first significant inhomogeneity visible at the resistivity profile RP8 interpretation (Mantlik et al., 2009) at the position 90–100 m from the beginning of the profile.

3.11.4 GRAVITY ANOMALY OF CHROMITE DEPOSIT BODY, CUBA: CASE STUDY

The geophysical techniques in mineral investigation depend on the variability of physical properties, as well as on local geological environment, topography, etc. The target of gravity interpretation is to locate and characterize the buried mineral source parameters—in particular, the density contrast, depth, and shape (Essa, 2014; Biswas, 2015; Mehanee and Essa, 2015). Figure 3.35 shows the gravity anomaly of length 180 m over a chromite (density 4.8 gm/cc) deposit body in the chromite region of the Camaguey area, Cuba (an island nation in the Caribbean Sea) (Roy, 2001; Essa and Munsch, 2019).

These chromite deposits may exist in a complex geological environment associated with serpentinized peridotite and dunite with slight quantities of gabbro, troctolite, and anorthosite. Uplift and erosion have detached the overlying rocks from the serpentine, except in synclinal areas, the largest of which extends from Central Lugareno to Loma Yucatan. This gravity anomaly has been interpreted by utilizing different inversion methods, as demonstrated in Table 3.8. Table 3.8 demonstrates that the estimated chromite deposit body parameters, amplitude factor (A), depth (z), location of the body (d), and shape (q), by utilizing these approaches, have a plausible consensus on the depth of burial achieved upon conducting drilling.

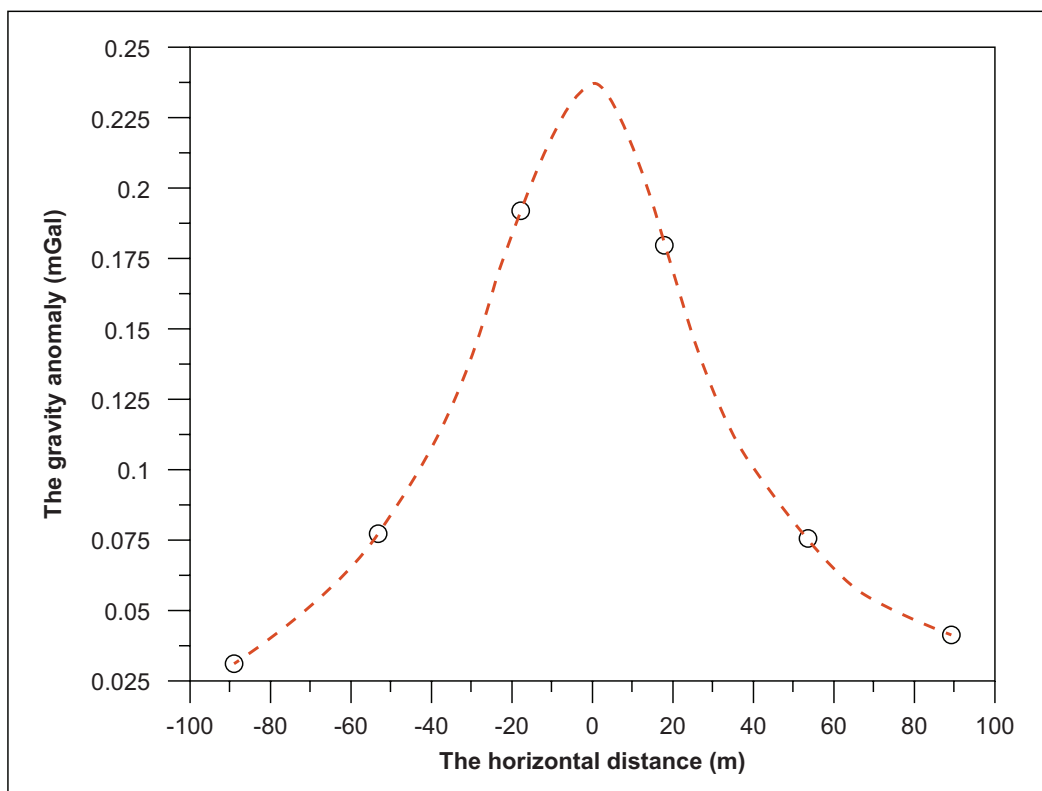


FIGURE 3.35 The gravity anomaly of chromite deposit body (Cuba).

TABLE 3.8**Results Obtained for Interpreting Gravity Anomaly of Chromite Deposit Body (Cuba) Using Different Inversion Methods**

Parameters	Drilling Information	Essa (2011) Method	Biswas (2015) Method	Ekinci et al. (2016) Method	Essa and Munsch (2019) Method
A (mGal m ²)	—	412.33	16.80	288.25	408.25
z (m)	21.00	21.02	42.30	23.23	21.15
d (m)	—	—	−2.40	58.73	0.63
q (dimensionless)	—	1.50	1.00	1.50	1.47

Source: Essa and Munsch (2019).

3.11.5 GRAVITY SURVEY IN DEVON AREA, ENGLAND: AN OUTCROPPING GEOLOGICAL CASE STUDY

The survey area (13×20 km²) is located in the extreme north-western corner of Devonshire in England (Alsadi, 1967; Alsadi and Baban, 2014). The latitude 51°05' N and the longitude 3°55' W form the southern and eastern boundaries of the area, respectively. The ground elevation lies in the range 275–300 m above sea level. There is a gradual decrease in elevation level in the western and south-western direction. The Middle Devonian consists of slates within which there are two 10 m bands of limestone. The Upper Devonian is dominantly made up of slates and sandstones that include the outcropping Pickwell Down sandstone member, which was mainly investigated in this case study. The succession of the Devonian rocks forms parallel tracts striking at a bearing of about 285°, with an average dip of 35° due south. The fold axes are generally pitching 15°–30° due west, for which the Bouguer anomaly contour map is presented by Alsadi and Baban (2014).

The geological structural model which is nearest to the realistic picture of the subsurface geology of the area is achieved through the interpretation of the final Bouguer gravity map, which remains one of the principal mottos of the gravity survey project. The input data to the interpretation process is the Bouguer anomaly map, which was drawn at 1 mGal intervals. There are qualitative and quantitative approaches of interpretation. After a descriptive approach and after trying to associate the main apparent changes with the available geological knowledge of the area, a more deterministic approach is followed. This includes computations of the regional and residual gravity, then gravity-to-geology translation using model analysis technique.

A visual inspection of this map revealed that the anomalies show a fairly constant strike, which follows the trend of the Devonian outcrop very closely. In addition, there exists a central region of anomaly disturbance. In general, the gravity values are noted to be falling in a northerly direction. The Bouguer contour map (Alsadi and Baban, 2014) shows a unique resemblance between the gravity picture and that of the known geological structure of the area. There is a close corroboration between the central gravity low and the outcrop of one of the Upper Devonian sedimentary members. Based on the general contour appearance and on the known geological structure of the area, the gravity map was considered as reflecting a two-dimensional geological model consisting of parallel rock layers dipping in a southerly direction. For this reason, interpretation was based on drawing gravity profiles perpendicular to strike of the geological layers. Due to the simplicity of the Bouguer gravity map, a manual method was used in establishing the regional anomaly. Based on the analysis of ten profiles taken at right angles to the general trend, it was found that the mean regional gradient is 1.2 ± 0.20 mGal per 1.6 km. For a density contrast of -0.15 gm/cc, these sediments must reach a thickness of about 5 km at the northern coast of Devon to account for the 30 mGal gravity change over the area.

The separated residual anomaly profiles were subjected to model analysis to ascertain the fact that anomalies caused by the dipping belt of sandstone and the dip angle of the formation decrease

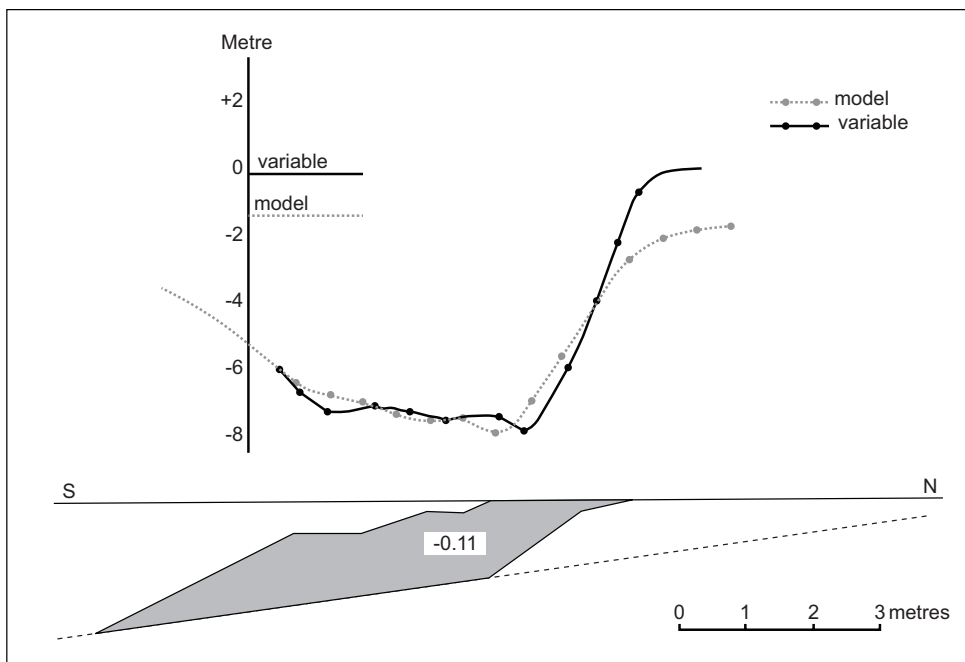


FIGURE 3.36 Gravity profile AB compared with the calculated anomaly for a theoretical model based on the thrust hypothesis model analysis.

with depth whilst, at a depth of about 3 km, the formation becomes nearly horizontal (Alsadi and Baban, 2014), as shown in Figure 3.36.

3.11.6 POSSIBLE TRIASSIC FOLDING BELOW ABU RASSAIN, THE WESTERN DESERT, IRAQ: A CASE STUDY

The area of study is situated in the west of Iraq, to the west of the town of Hatra (Alsadi and Baban, 2014). Previously, the area was covered by gravity surveys (Sayyab and Valek, 1968; Alsaikh and Baban, 1991) and by aeromagnetic surveys in 1978. A detailed seismic reflection survey was made over the Western Desert, Iraq (Mohammed, 1981), and the study produced a topographical map of the basement and an isopach map of the Lower Palaeozoic rocks. In the following analysis, the gravity features of local nature observed on the gravity map will be concentrated upon. The Bouguer anomaly map was prepared by the Iraqi Petroleum Company (IPC), with a contour interval of 1 mGal used (Figure 3.37).

It is dominated by a large oval positive anomaly occurring in the middle of the area. It has a sharp northern gradient, with the well Kh-1 lying in the middle of it. The southern and eastern gradients are gentler. Other local disturbances can also be seen. The smooth regional map shows the oval gravity high in the middle having a sharp gradient towards the north (0.6 mGal/km), and it has maximum amplitude of 20 mGal (Alsadi and Baban, 2014). The residual map (Figure 3.38) shows a number of local anomalies, of which the anomaly (I), referred to here as the Abu Rassain anomaly, is the most prominent, and it further extended within the Khleisia graben (Ditmar, 1971). Ditmar's evidence for the graben is apparently the linear irregularities of the contours (Figure 3.37), which stretch from Hatra south-westwards without information about its deep origin (Mohammed, 1981). To the north-east of Abu Rassain (III), a positive anomaly occurs, another large, three-dimensional one, referred to here as the Abu Jurd anomaly. No information is available regarding the direct and detailed density measurements of the rocks involved in the Western Desert. However, Ditmar

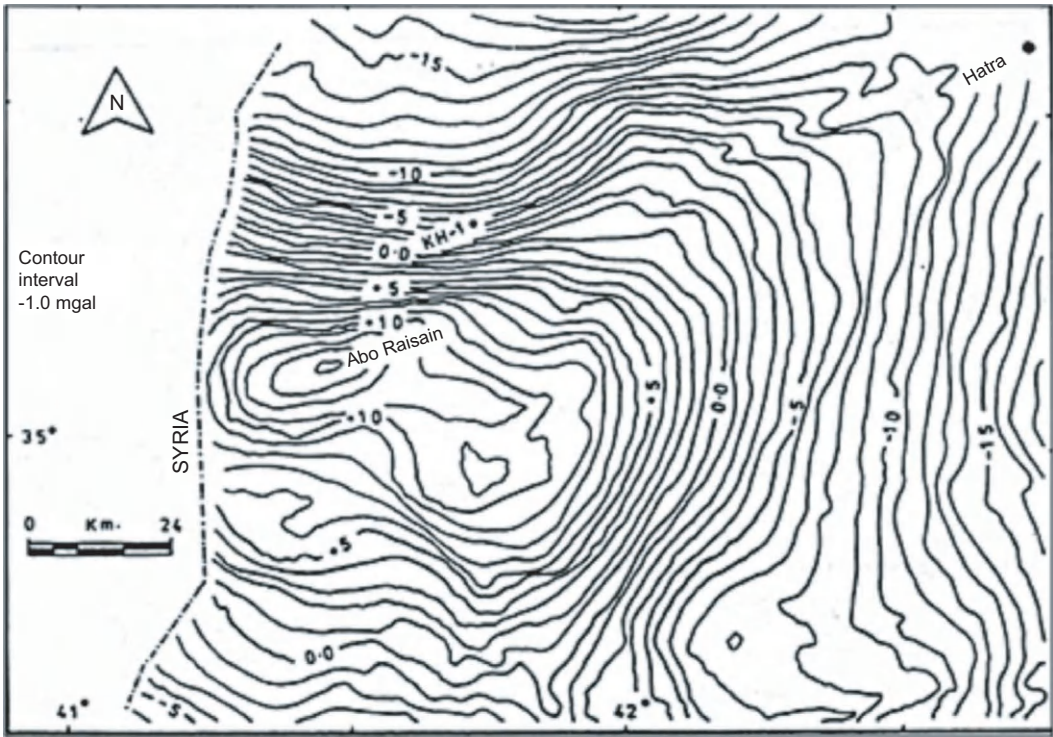


FIGURE 3.37 Bouguer anomaly map of the area.

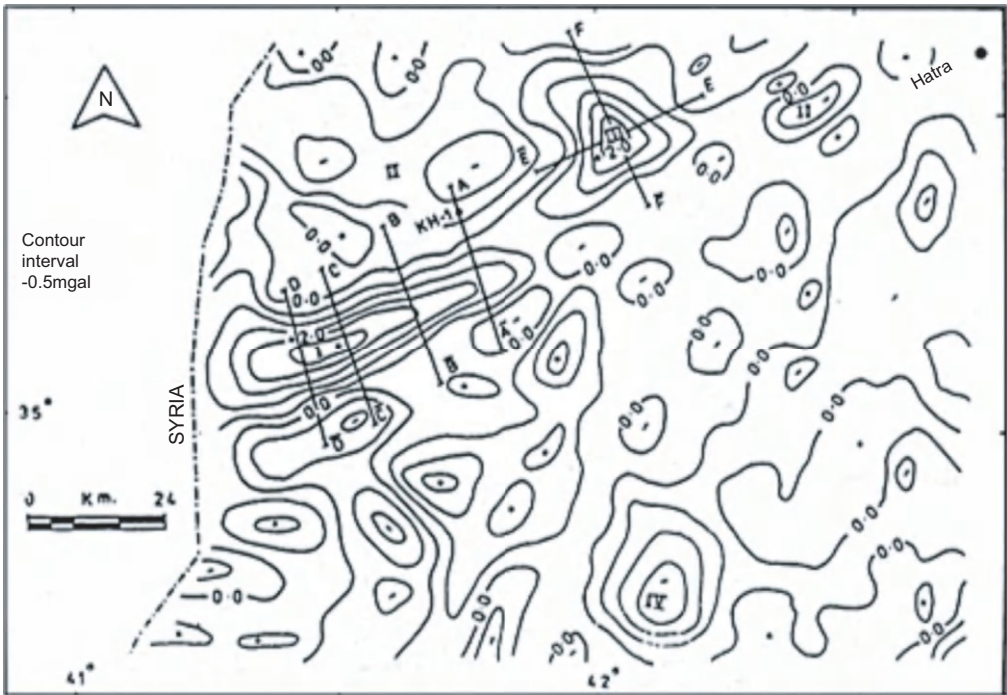


FIGURE 3.38 Residual anomaly map of the area, and AA, BB, CC, DD, EE, and FF are the chosen profile locations for modelling purposes.

TABLE 3.9**Average Density of the Systems of the Geological Column in the Western Desert**

System	Average Density (gm/cc)
Middle and Lower Miocene	2.64
Palaeogene to Middle Cretaceous	2.46
Lower Cretaceous	2.61
Jurassic to Upper Triassic	2.71
Middle Triassic to Cambrian	2.62

(1971) gave the average density of the systems of the geological column in the Western Desert, as given in Table 3.9.

To locate the source of the Abu Rassain and Abu Jurd anomalies, a density contrast of 0.18 gm/cc is considered representative for the difference between the Triassic (Kurra Chine Formation) and the later successions (Ditmar, 1971; Mohammed, 1981). Obviously, their contour maps have not included inclination due to the smoothing procedure of the observed contours. This slight extra regional effect is subtracted from the observed curve to yield a simple high over the uplift. The best cross section of the uplift was found by trial and error using a computer program to compute the theoretical values of gravity of the two-dimensional models.

The suggested model is an elongated uplift whose topmost rocks are made up of Upper Triassic. It is worth mentioning that the uplift is not symmetrical, with the northern limb having greater inclination. It is possible that the northern limb is faulted, hence becoming steeper. Alsadi and Baban (2014) shows a subsurface contour map on top of the Upper Triassic as deduced from the aforementioned considerations.

Abu Jurd, a three-dimensional positive anomaly with maximum amplitude of 2.4 mGal, lies to the north-east of well Kh-1 (Figure 3.37). It is believed that the gravity anomaly may be associated with the uplifting of the unconformity above the Triassic rocks. To obtain the geological structure causing the anomaly, two profiles (FF, EE) running normal to each other are considered (Figure 3.38). A similar procedure as that used for the Abu Rassain anomaly, which is followed by the computer program, was used for calculating the gravitation effect of a three-dimensional model (Alsadi and Baban, 2014). The interpretation suggested the solution for a nearly circular uplift of the Upper Triassic, which is slightly asymmetrical, with its top occurring at a depth of 530 m. A subsurface structure contour map of the uplift is shown in Figure 3.39.

3.11.7 APPLICATION OF THE GRAVITY METHOD IN ARCHAEOLOGICAL PROSPECTION IN JORDAN: A CASE STUDY

In comparison to other geophysical methods, the gravity method has seen limited use, mainly because of the deployment of more human resources and the enormous time and energy involved for adequate elevation and terrain corrections (Linnington, 1966; Fajkiewicz et al., 1982). A micro-gravimetric and gravity gradient methods were used for engineering applications and to search for underground cavities (Arzi, 1975; Blizkovsky, 1979; Butler, 1984b; Seguin et al., 1994). The application of an integrated geophysical method involving the **gravity** method in the Raniganj Coalfield of India identified and delineated the abandoned, unknown, unsurveyed coal workings (Seguin et al., 1994, 1995). A similar study was made to outline the underground structures at the Umm er-Rasas archaeological site in Jordan (Batayneh et al., 2006).

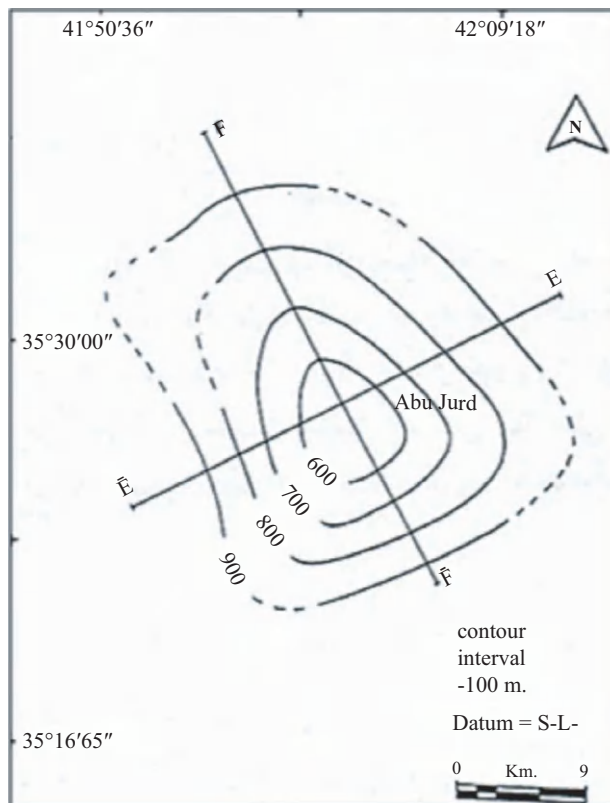


FIGURE 3.39 Structural contour map on top of the Upper Triassic (Abu Jurd anomaly).

3.12 GRAVITY DATA OF INDIA FOR DELINEATION OF STRUCTURAL FABRIC AND CRUSTAL ARCHITECTURE

In recent years, a large number of gravity data have been recorded in the different parts of India by several institutions for their different projects. These data were compiled in the 70s, and a map of 10 mGal ($1 \text{ mGal} = 10^{-3} \text{ cm/sec}^2$) interval on 1:5 million scale was prepared. However, such a map is not adequate enough for present-day exploration requirements and was mostly utilized only for regional lithospheric and geodynamic studies (Tiwari et al., 2013). In order to improve upon the gravity anomaly maps prepared in the late 70s and in the 80s, an effort was made to bring five major research/government organizations, namely, the CSIR National Geophysical Research Institute (CSIR-NGRI), Survey of India (SOI), Geological Survey of India (GSI), Oil and Natural Gas Corporation (ONGC), and Oil India Ltd. (OIL), to collaborate with the primary goal of updating and publishing a gravity anomaly map of India from the basic gravity data collected under various projects by these five collaborative organizations. A virtual databank combining data from all the organizations for the entire India, totalling 143,786, is archived, and the grid was prepared at GSI, Hyderabad. For uniform coverage over the entire country, an optimal spatial interval of 3 minutes' arc interval is chosen, thereby only 51,356 gravity observations are considered to prepare the revised gravity anomaly map of India 2006, with a contour interval of 5 mGal (Chapter 14; Figure 14.2–3). Since data were heterogeneous in quality, the type of corrections and their absolute value determinations were always not based on the International Gravity Standardization Network (IGSN-1971) (Morelli et al., 1971). There were extra efforts in refining the gravity data by reprocessing, by incorporating detailed terrain corrections. The corrected data were replotted, and the

Bouguer anomaly maps were printed on the 1:2 million scale (Sundaram et al., 2007). The Bouguer maps were superimposed on the geological map published by the GSI on the same scale, which reveal a one-to-one correlation between the number of anomalies and the geological structure/mineralized zone. The availability of gravity anomaly maps provides vital information for planning and studies ranging from mineral/hydrocarbon exploration to tectonic studies (Mishra et al., 2008). These maps can be procured from the GSI, details of which are stated at webpages (<http://ngri.org.in/pdffiles/aboutngri/gravitymap.pdf>; www.gsi.gov). One of these maps (GMSI, 2006), the Bouguer gravity anomaly map of India (image), is presented in Figure 3.40.

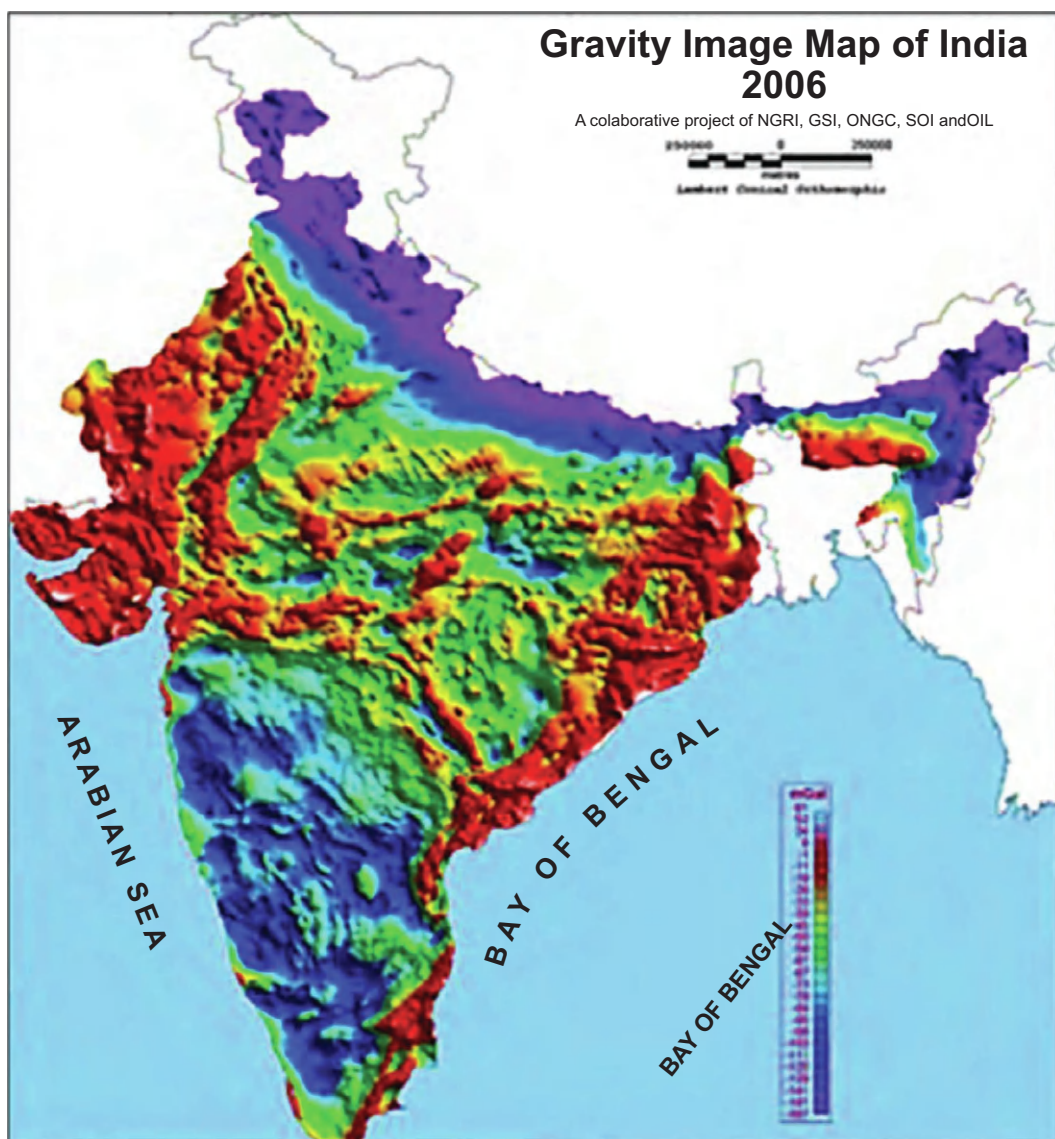


FIGURE 3.40 The Bouguer gravity anomaly map of India, depicting the nature of the anomalies in the different parts of India. High (red) and low (blue) values of gravity anomalies indicate high and low subsurface density distributions, respectively.

3.13 INTERPRETATION OF GRAVITY MAPS AND GRAVITY SIGNATURES OF MINERAL DEPOSITS: A REVIEW

3.13.1 GRAVITY MAPS OF MINERAL DEPOSITS

The first step in the interpretation is to make preliminary assessment of processed data without any geological bias. The reduced or corrected gravity data is plotted in the form of profiles/contours, or both, with appropriate scale (Figure 3.41). Such profiles or maps are carefully examined to identify anomalous bodies and their nature. Contour patterns provide valuable information of subsurface material. The identification of low/high, gravity gradient, and symmetry patterns is done to assess the shape and dimensions of the subsurface structure or body. It is understood that the magnitude of gravity anomalies is mostly related to the density contrast (change in density) across the map area. The anomaly wavelength (broadness, horizontal gravity gradient) is related to the depth of the lateral change in density (Figure 3.42).

From the gravity anomaly maps, the following geological inferences can be made (Figure 3.43). These should be considered as a guideline, not as a rule:

1. *Circular anomalies.* Compact mineral body or salt dome.
2. *Elongated anomalies.* Graben/buried folds/buried channels.
3. *Negative anomalies.* Weaker or less-dense rocks, such as sedimentary basins/batholiths.
4. *Positive anomalies.* More dense rocks, such as ultramafics/mafic.
5. *Wavelength of anomalies.* The deeper the body, the broader and lower the amplitude of the profile.
6. *Gradient and curvature.* A rapid change in amplitude/gradient suggests density change in the subsurface, such as faults or the edge of a buried basin.

Some geological structures that can be approximated by regular geometrical shapes follow (Table 3.10).

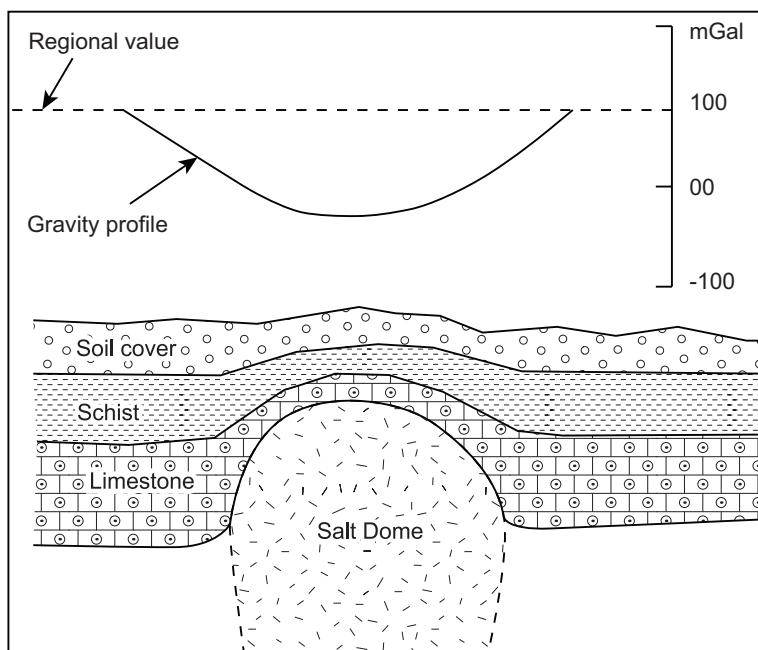


FIGURE 3.41 Probable gravity profile along with the geological environment for interpreting gravity data.

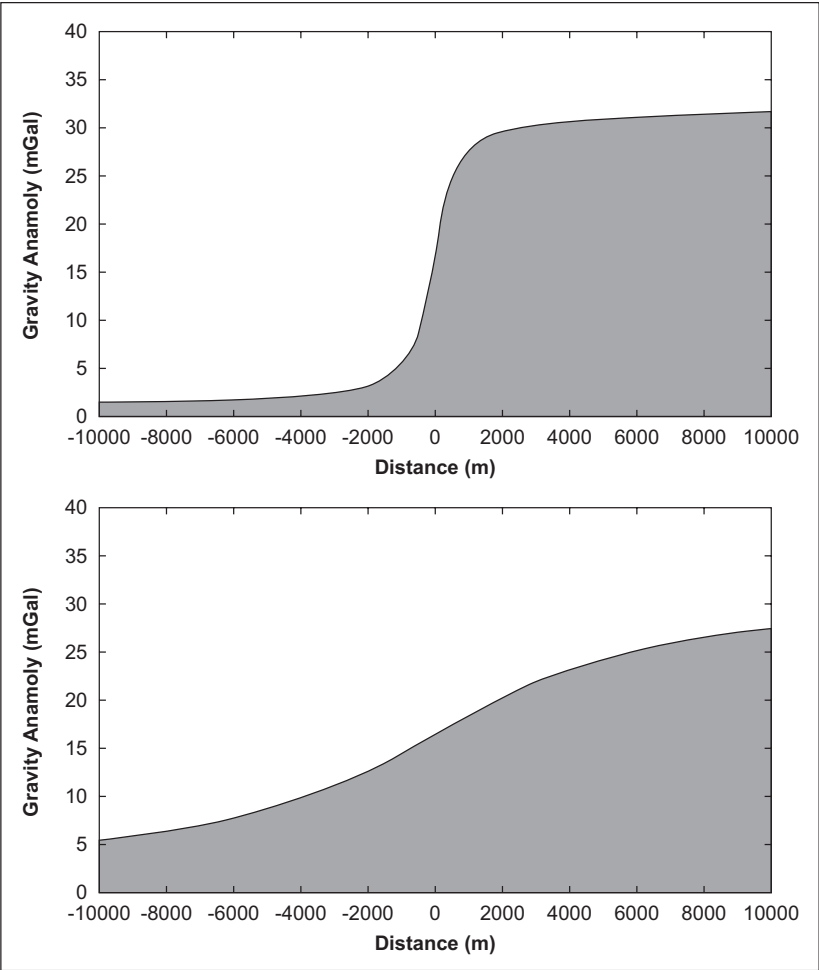


FIGURE 3.42 Anomaly wavelengths of the causative sources in order to interpret gravity data.

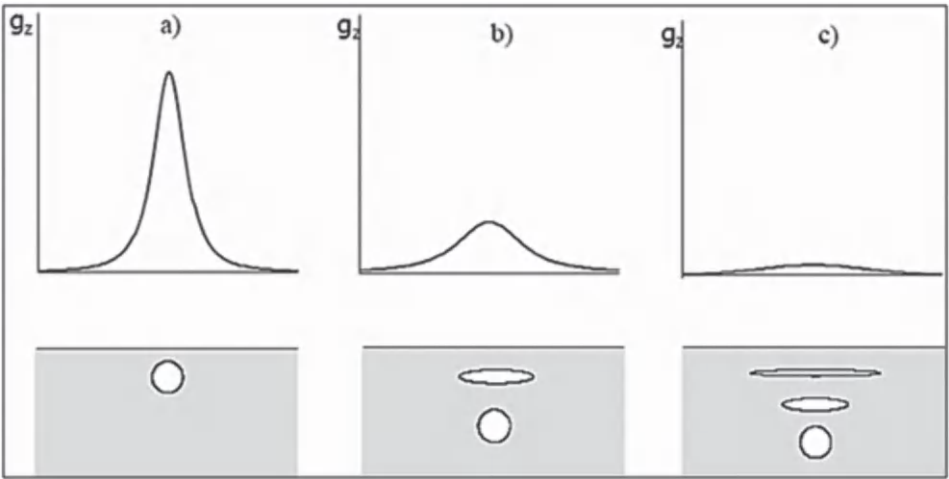


FIGURE 3.43 Gravity responses show the change in gradient/amplitude depending on the subsurface geological sources.

TABLE 3.10**Geometrical Shape Corresponding to Geological Structures**

Geometrical Shape	Geological Approximation (Ore Bodies/Geological Structures)
Sphere	Highly compacted ore deposits
Horizontal cylinder	Anticline/synclines and mine cuts
Vertical cylinder	Kimberlite pipes
Thin horizontal plate	Conrad/Moho discontinuity
Thin vertical/dipping plate	Fractures, mineralized veins, shear zone
Thin vertical/dipping plate	Dykes and other intrusive body

Note: These are not unique solutions. Several models can produce exactly the same anomaly. Therefore, geological understanding is necessary to limit the possible solutions.

Source: Ramakrishna (2006).

3.13.1.1 Regional and Residual Anomaly

As discussed in beginning, a gravity anomaly originates from the inhomogeneous distribution of density in the Earth. Suppose that the density of rocks in a subsurface body is ρ , and the density of the rocks surrounding the body is ρ_0 . The difference $\Delta\rho = \rho - \rho_0$ is called the density contrast of the body with respect to the surrounding rocks. If the body has a higher density than the host rock, it has a positive density contrast; a body with lower density than the host rock has a negative density contrast.

3.13.1.1.1 Estimation of Depth of Anomalous Mass

Gravity anomaly is affected by the dimensions of the anomalous mass, the density contrast between the host and the anomalous mass, and the depth of the anomalous body. We generally infer that the horizontal extent of an anomaly represents wavelength, whilst the wavelength of an anomaly is a measure of the depth of the anomalous mass. It is therefore the large, deep bodies that give rise to broad (long-wavelength), low-amplitude anomalies, while small, shallow bodies cause narrow (short-wavelength), sharp anomalies.

It is observed that Bouguer gravity anomalies contain superposed anomalies from several sources. The long-wavelength anomalies due to deep-density contrasts are called ‘regional anomalies’. They are important for understanding the large-scale structure of the Earth’s crust under major geographic features, such as mountain ranges, oceanic ridges, and subduction zones. In contrary to large-scale anomaly, short-wavelength anomalies are ‘residual anomalies’; these are due to shallow anomalous masses that may be of commercial interest. One has sufficient geological knowledge for the area for interpreting the residual anomalies.

3.13.1.2 Separation of Regional and Residual Anomalies

The separation of anomalies of regional and local origin is an important step in the interpretation of a gravity map (Figure 3.44a and b). This may be performed graphically by sketching a linear or curvilinear line (based on visual estimation). Such a method is biased by the interpreter, but this is not necessarily disadvantageous, as geological knowledge can be incorporated into the selection of the regional field. Such analysis may be based on selected profiles across some structure. There are numerous techniques ranging from simple visual inspection to advanced mathematical analysis:

1. Graphical method/smoothing method
2. Grid method
3. The least squares method (polynomial or trend surface analysis)
4. Fourier series
5. Anomaly enhancement and wavelength filtering (use of low-pass filters and high-pass filters)

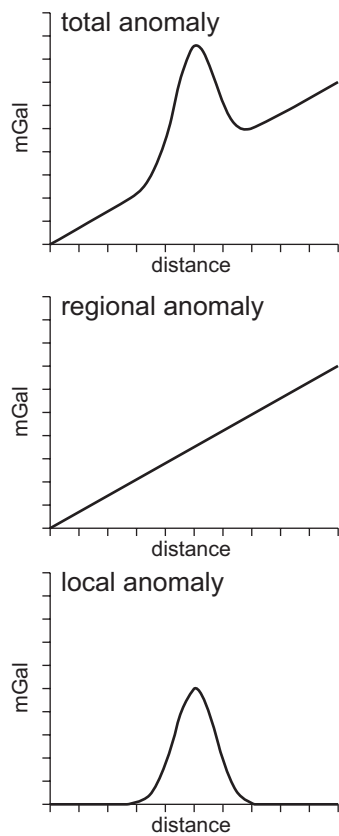


FIGURE 3.44 (A) The Bouguer gravity anomalies as total, regional, and local subsurface geological sources.

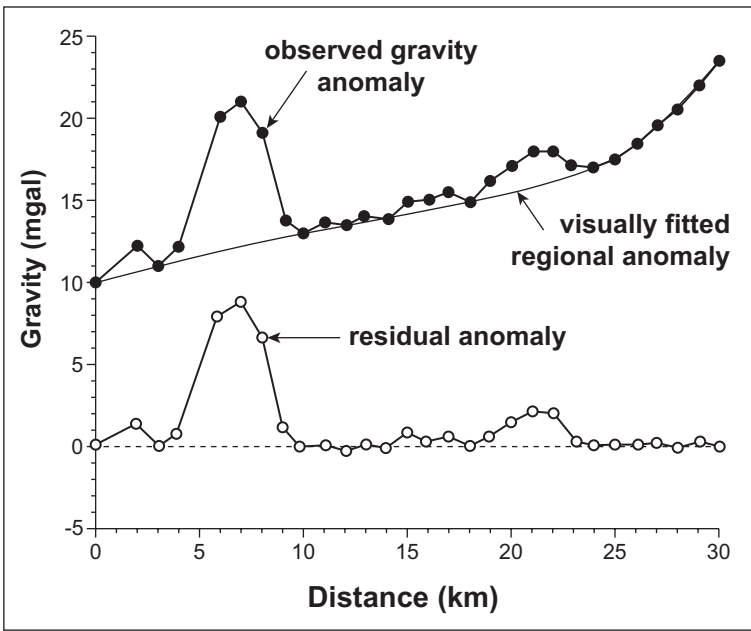


FIGURE 3.44 (B) The regional and residual separation of the Bouguer gravity anomalies.

3.13.2 GRAVITY SIGNATURES OF MINERAL DEPOSITS

A mineral deposit, as an anomalous unit of metalliferous minerals, contains minerals with quite different physical properties to those of country rocks. Pyrite, pyrrhotite, and magnetite are some common minerals in ore deposits, all of which have distinctive physical properties and may greatly affect the geophysical response (Hinze et al., 2013).

1. *Ore compositions.* It is very much known the fact that the composition of the ore deposit and the associated minerals has control of the density of the body, which depends on the elementary composition of minerals. Several metals have very high specific densities. The geophysical surveys also revealed that the density of sulphide minerals was found to be very abnormally high, in the range of generally greater than 4.0 gm/cc, whilst the mafic and ultramafic rocks hosting Ni-Cu-PGE deposits also exhibit high densities because of the abundance of mafic minerals such as olivine and pyroxene. Thus, density may be used for the direct detection and quantitative measurement of many sulphide ores.
2. *Form of the ore bodies.* The form of the ore bodies, such as the size, shape, orientation, and depth. Ore mineral distribution and texture are very important to understand their significant roles in generating gravity anomalies. It is observed that the size, orientation, and depth extent of mineral deposits are the main factors with regard to geophysical expressions. It has been discussed earlier in this chapter that the causative bodies at deeper layers suppress geophysical signatures. It has been established that the gravity and magnetic methods only detect and delineate the lateral contrasts in density or magnetization, whilst the electrical and seismic methods can detect vertical, as well as lateral, contrasts of the resistivity and velocity or reflectivity.
3. *Geological structures.* It is observed that most of the mineral deposits are structurally controlled, while the mineral occurrences are often found restricted to structural elements, such as faults, shear zones, and lithological unconformities; some deposits form stratiform bodies, while others are formed in a specific stratigraphic interval as a stratabound formation.

Table 3.11 provides the applicability of different geophysical methods in the exploration of some minerals' deposits.

TABLE 3.11
Different Geophysical Methods for Mineral Targeting

GP	*Generally Ineffective		**Moderately Effective				***Highly Effective				
	Air or Ground	Application	Ni-Cu-PGE	Fe-Ti BIF	Gold	VMS	Olympic Dam Type	SEDEX	Porphyry Cu	Pb-Zn	Diamonds
Magnetic	Air	Geological framework	***	***	***	***	***	***	***	***	***
		Direct targeting	**	***	*	***	***	**	***	*	***
	Ground	Geological framework	**	***	***	***	***	***	***	**	***
		Direct targeting	**	***	**	***	***	**	***	*	***

(Continued)

TABLE 3.11 (Continued)**Different Geophysical Methods for Mineral Targeting**

*Generally Ineffective			**Moderately Effective				***Highly Effective				
GP	Air or Ground	Application	Ni-Cu-PGE	Fe-Ti BIF	Gold	VMS	Olympic Dam Type	SEDEX	Porphyry Cu	Pb-Zn	Diamonds
EM	Air	Geological framework	***	***	**	**	*	**	*	*	**
		Direct targeting	***	***	**	***	**	**	*	*	***
	Ground	Geological framework	***	***	**	**	*	**	*	**	***
		Direct targeting	***	**	*	***	**	**	*	**	***
Electrical	Ground	Geological framework	***	*	*	**	**	*	*	**	*
		Direct targeting	***	**	**	***	**	*	***	***	**
Gravity	Air	Geological framework	***	***	**	**	**	**	**	**	**
		Direct targeting	***	***	*	***	**	**	*	**	***
	Ground	Geological framework	***	***	**	**	**	**	**	**	***
		Direct targeting	***	***	*	***	**	***	**	***	***
Radiometric	Air	Geological framework	**	**	***	***	***	***	***	**	**
		Direct targeting	*	*	**	**	***	*	***	*	*
	Ground	Geological framework	**	*	***	**	***	**	***	*	*
		Direct targeting	*	**	**	**	***	*	***	*	*
Seismic	Ground	Geological framework	**	**	**	**	**	**	**	**	**
		Direct targeting	**	**	*	**	**	**	**	**	**

Note: GP, geophysical method; EM, electromagnetic; PGE, platinum group of elements; BIF, banded iron formation; VMS, volcanogenic massive sulphide; SEDEX, sedimentary exhalative deposits.

Source: Airo (2015); Ramakrishna (2006).

REFERENCES

- Airo, M.-L., 2015. Geophysical signatures of mineral deposit types in Finland. *Geological Survey of Finland, Special Paper*, vol. 58, p. 144.
- Alsadi, H.N., 1967. A gravity investigation of the Pickwell Down Sandstone, North Devon. *Geological Magazine*, vol. 104, no. 1, pp. 63–72.
- Alsadi, H.N. and Baban, E.N., 2014. Introduction to gravity exploration method. Book, First Edition, University of Sulaimaniyah, Kurdistan Region, Iraq. Copyright 2014 by Hamid N. Alsadi and Ezzadin N. Baban. Printed by Paiwand press, July 2014, Sulaimaniyah, Kurdistan Region, Iraq.

- Alsaikh, Z.D. and Baban, E.N., 1991. Possible Triassic folding below Abu Rassain, the Western Desert, Iraq. *Journal of Natural Science*, vol. 1, no. 1A, pp. 28–34.
- Arzi, A., 1975. Microgravimetry for engineering applications. *Geophysical Prospecting*, vol. 23, pp. 408–425.
- Batayneh, A., Khataibeh, J., AlRshdan, H., Tobasi, U. and Al-Jahed, N., 2006. The use of microgravity, magnetometry and resistivity surveys for the characterization and preservation of an archaeological site of Umm er-Rasas, Jordan. *Archaeological Prospection*, vol. 14, pp. 60–70.
- Beltrao, J.F., Silva, J.B.C. and Costa, J.C., 1991. Robust polynomial fitting method for regional gravity estimation. *Geophysics*, vol. 56, pp. 80–89.
- Benson, R., Glaccum, R. and Noel, M., 1983. Geophysical techniques for sensing buried waste and waste migration. Environ. Monitor. Syst. Lab., Off. Res. Develop., US Environ. Protect. Ag., Las Vegas, NV, Rep. 68–03–3050, p. 114.
- Birch, F.S., 1961. The velocity of compressional waves in rocks to 10 kilobars. *Journal Geophysical Research*, vol. 66, pp. 2199–2224.
- Biswas, A., 2015. Interpretation of residual gravity anomaly caused by a simple shaped body using very fast simulated annealing global optimization. *Geoscience Frontiers*, vol. 6, pp. 875–893.
- Biswas, U.K., Ghatak, T.K., Ghosh, D.C., Naskar, D.C., Chakraborty, P.K. and Choudhury, K., 1999. Application of geoelectrical survey for detection of voids in Raniganj coalfield, W. B. GSI Spl. Pub. (Golden Jubilee Seminar on Exploration Geophysics in India), no. 49, pp. 471–480.
- Biswas, U.K., Ghosh, D.C., Naskar, D.C. and Chakraborty, P.K., 2005. Detection of subsurface voids by surface geophysical methods in Baktarnagar area of Raniganj coalfield, W.B. Proceedings of Workshop on Solid Fossil Fuel. An introspection and perception in Indian context. Geological Survey of India Spl. Publ. 81, pp. 198–210.
- Blizkovsky, M., 1979. Processing and application in microgravity surveys. *Geophysical Prospecting*, vol. 27, pp. 848–861.
- Butler, D.K., 1984a. Gravity gradient determination concepts. *Geophysics*, vol. 49, pp. 828–832.
- Butler, D.K., 1984b. Micro-gravimetric and gravity gradient techniques for detection of subsurface cavities. *Geophysics*, vol. 49, pp. 1084–1096.
- Cassinis G., 1930. Sur l'adoption d'une formule Internationale pour la p'ésanteur normale. *Bull. Geod.*, vol. 26, pp. 40–49.
- Ditmar, V., 1971. *Geological Conditions and Hydrocarbons Prospects of the Republic of Iraq*. Baghdad: OEC Library, unpublished.
- Dobrin, M.B., 1976. *Introduction to Geophysical Prospecting*. 3rd edn. New York: McGraw-Hill.
- Ekinci, Y., Balkaya, C., Gokturkler, G. and Turan, S., 2016. Model parameter estimations from residual gravity anomalies due to simple-shaped sources using differential evolution algorithm. *Journal of Applied Geophysics*, vol. 129, pp. 133–147.
- Elkins, T.A., 1951. The second derivative method of gravity interpretation. *Geophysics*, vol. 16, pp. 29–50.
- Essa, K.S., 2011. A new algorithm for gravity or self-potential data interpretation. *Journal of Geophysics and Engineering*, vol. 8, pp. 434–446.
- Essa, K.S., 2014. New fast least-squares algorithm for estimating the best-fitting parameters of some geometric-structures to measured gravity anomalies. *Journal of Advanced Research*, vol. 5, pp. 57–65.
- Essa, K.S. and Munschy, M., 2019. Introductory chapter: Mineral exploration from the point of view of Geophysicists. In: *Minerals Book*, from the edited volume. London, UK, DOI: 10.5772/intechopen.84830.
- Fajkiewicz, Z.J., 1976. Gravity vertical gradient measurements for the detection of small geologic and anthropomorphic forms. *Geophysics*, vol. 41, pp. 1016–1030.
- Fajkiewicz, A., Glinski, A. and Sliz, J., 1982. Some applications of the underground tower gravity vertical gradient. *Geophysics*, vol. 47, pp. 1688–1692.
- Foulger, G.R. and Pierce, C. 2014. Geophysical methods in geology. Available at www.dur.ac.uk/g.r.foulger/Teaching/GG_HandoutsAll.pdf, March 15, 2014.
- Galland, P. and Stierman, D., 2005. Gravity investigation of the King Road landfill, Lucas County, Ohio North-Central Section -39th Annual Meeting, Geological Society of America Minneapolis, MN, pp. paper 46–4.
- GMSI (Gravity Map Series of India), 2006. Publisher National Geophysical Research Institute & Geological Survey of India, Hyderabad, India. Available at <http://ngri.org.in/pdf/files/aboutngri/gravitymap.pdf>.
- Gokarn, S.G., Rao, C.K., Gupta, G., Singh, B.P. and Yamashila, M., 2001. Deep crustal structure in Central India using magnetotelluric studies. *Geophysical Journal International*, vol. 144, pp. 685–694.
- Greenhouse, J., Monier-Williams, M., Ellert, N., and Slaine, D., 1989. Geophysical methods in water contamination studies. Proceedings of Exploration '87, Ont. Geo. Sur. Spec., vol. 3, pp. 666–677.

- GSI, 2016. Institutional Report. *Proc. Indian Natn. Sci. Acad.*, vol. 82, no. 3 July Spl Issue 2016 pp. 1061–1081. Printed in India, DOI: 10.16943/ptinsa/2016/48504.
- Hammer, S., 1939. Terrain correction for gravimeter stations. *Geophysics*, vol. 4, pp. 184–89.
- He, H., Fang, J., Guo, D., Cui, R. and Xue, Z., 2024. 3D density imaging using gravity and gravity gradient in the wavenumber domain and its application in the Decorah. *Scientific Reports*, vol. 14, no. 134(2024), p. 14.
- Heiskanen, W.A. and Moritz, H., 1967. *Physical Geodesy*. San Francisco: Freeman.
- Hinze, W.J., 1990. The role of gravity and magnetic methods in engineering and environmental studies. In: Ward, S.H. (ed), *Geotechnical and Environmental Geophysics*, vol. 1. Tulsa, OK: Society Exploration Geophysicists, pp. 75–126.
- Hinze, W.J., 2013. *Gravity Method*. New York: Cambridge University Press, pp. 19–37.
- Hinze, W.J., von Frese, R.R.B. and Saad, A.H., 2013. *Gravity and Magnetic Exploration: Principles, Practices and Applications*. From Part I. New York: Cambridge University Press, pp. 19–37.
- Hood, P., 1965. Gradient measurement in aeromagnetic surveying. *Geophysics*, vol. 30, pp. 891–902.
- International Association of Geodesy, 1980. The geodesist's handbook. Ed. I.I. Mueller. *Bulletin Géodésique*, vol. 54, no. 3.
- James, W.R., 1966. Fortran IV program using double Fourier series for surface fitting of irregularly spaced data. In *Kansas Geological Survey Computer Contribution*, p. 5.
- Johnson, G.R. and Olhoeft, G.R., 1984. Density of rocks and minerals. In Carmicheal, R.S. (ed) *Handbook of Physical Properties of Rocks*, vol. 3. Boca Raton, FL: CRC Press, pp. 1–38.
- Kaila, K.L., Reddy, P.R., Dixit, M.M. and Koteswara Rao, P., 1985. Crustal structure across the Narmada-Son lineament Central India from deep seismic sounding. *Journal of Geological Society of India*, vol. 26, pp. 465–480.
- Kearey, P., Brooks, M. and Hill, I., 2002. *An Introduction to Geophysical Exploration*. 3rd edn. Malden, MA: Blackwell Science, ISBN 0632049294.
- Keating, P.B., 1998. Weighted Euler deconvolution of gravity data. *Geophysics*, vol. 63, pp. 1595–1603.
- Lerch, F.J., Klosko, S.M., Laubscher, R.E. and Wagner, C.A., 1979. Gravity model improvement using Geos 3 (GEM 8 and 10). *Journal of Geophysical Research*, vol. 84, pp. 3897–3916.
- Linington, R., 1966. Test use of a gravimeter on Etruscan chamber tombs at Cerveteri. *Prospezioni Archeologiche*, vol. 1, pp. 37–41.
- Lowrie, W., 2007. *Fundamentals of Geophysics*. 2nd edn. Published in the United States of America by New York: Cambridge University Press.
- Mantlik, F., Matias, M., Lourenco, J., Grangeia, C. and Tareco, H., 2009. The use of gravity methods in the internal characterization of landfills—A case study. *Journal of Geophysics and Engineering*, vol. 6, no. 4, pp. 357–364.
- Mariita, N.O., 2009. The gravity method. Kenya Electricity Generating Company Ltd-KenGen P.O. Box 785, Naivasha KENYA nmariita@kengen.co.ke. Presented at Short Course IV on Exploration for Geothermal Resources, organized by UNU-GTP, KenGen and GDC, at Lake Naivasha, Kenya, November 1–22, 2009.
- Matias, M., Silva, M., Ferreira, P. and Ramalho, E., 1994. A geophysical and hydrogeological study of aquifers contamination by a landfill. *Journal of Applied Geophysics*, vol. 32, pp. 155–162.
- Mehanee, S.A. and Essa, K.S., 2015. 2.5D regularized inversion for the interpretation of residual gravity data by a dipping thin sheet: Numerical examples and case studies with an insight on sensitivity and non-uniqueness. *Earth, Planets and Space*, vol. 67, p. 130.
- Mickus, K.L., 1980 (uploaded on November 2014). Gravity method: Environmental and engineering applications. *Geophysics*.
- Mishra, D.C., Tiwari, V.M. and Singh, B., 2008. Geological significance of gravity studies in India. Golden Jubilee Memoir 66. *Journal of the Geological Society of India*, pp. 329–372.
- Mohammed, A.M., 1981. Geophysical study to establish basement configuration in the western desert. M.Sc. Thesis, Univ. of Mosul., Unpublished.
- Morelli, C., Gantar, C., Honkasalo, T., McConnell, R.K., Tanner, J.G., Szabo, B., Uotila, U. and Whalen, C.T., 1971. The International Gravity Standardization Net 1971 (I.G.S.N.71) prepared for Air Force Cambridge Research Laboratories European Office of Aerospace Research and Development. Performing Organization Name and Address: Osservatorio Geofisico Sperimentale Viale R. Gessi 4 34123 Trieste, Italy. Controlling Office Name and Address Air Force Cambridge Research Laboratories/LW L.G. Hanscom AFB, Bedford, MA 01730, Spl. Pub. No. 4.

- Moritz, H., 1980. Geodetic reference system 1980. *Bulletin G  od  sique*, vol. 54, pp. 395–405.
- Mushayandebvu, M.F., Lesur, V., Reid, A.B. and Fairhead, J.D., 2004. Grid Euler deconvolution with constraints for 2D structures. *Geophysics*, vol. 69, pp. 489–496.
- Nafe, J. and Drake, C., 1957. Variation with depth in shallow and deep water marine sediments of porosity, density and the velocities of compressional and shear waves. *Geophysics*, vol. 22, pp. 523–552.
- Naskar, D.C., 2016. Tectonic and crustal understanding of lesser Himalayas along Purnea-Sevoke transect through integrated geophysical studies. *Bollettino di geophysica teorica ed applicata*, vol. 57, no. 4, pp. 301–312.
- Naskar, D.C., Das, L.K., Majumdar, R.K., Choudhury, K. and Roy, K.K., 2008a. Delineation of structural fabric and crustal architectures by gravity interpretation in Central India. *Indian Journal of Geology*, vol. 77, nos. 1–4, pp. 45–53.
- Naskar, D.C., Das, L.K., Roy, K.K. and Singh, O.P., 2008b. Structural studies of crust and upper mantle in parts of Madhya Pradesh, Central India. *Geological Survey of India, Pan-African Event: India & Antarctica*, Spl. Publ. no. 91, pp. 211–228.
- Naskar, D.C., Chakraborty, P., Ramasatish, T. and Tripathi, A., 2020. Crustal structure from gravity and magnetic data in parts of Nalgonda and Khammam Districts, Telangana (India). *Journal Indian Geophysical Union (IGU)*, vol. 24, no. 2, pp. 1–6.
- Naskar, D.C. and Gyanesh Lashkari, 2019. Geophysical study in Tonk and Dausa districts, Rajasthan: Its implication in regional geology. *Journal of Indian Geophysical Union*, vol. 23, no. 6, pp. 485–498.
- Naskar, D.C., Majumdar, N., Ajay, K. and Nikhil, K., 2020. Geophysical mapping in parts of Nuapada, Bolangir and Kalahandi districts of Odisha (India). *Journal of Indian Geophysical Union*, vol. 24, no. 3, pp. 27–41.
- Naskar, D.C., Majumdar, R.K., Choudhury, K. and Das, L.K., 2003. Magnetic profiles across NSL in parts of Central India. *Indian Minerals*, vol. 57, nos. 1–2, pp. 53–60.
- Nettleton, L.L., 1939. Determination of density for reduction of gravimeter observations. *Geophysics*, vol. 4, pp. 176–183.
- Radogna, P.V., Olivier, R. and Logean, P., 2004. A Subway Project In Lausanne, Switzerland, As An Urban Microgravimetry Test Site: Acquisition, Correction Of Buildings Influence And Modeling. European Association of Geoscientists & Engineers, Source: Conference Proceedings, 17th EEGS Symposium on the Application of Geophysics to Engineering and Environmental Problems, Feb 2004, cp-186–00088, DOI: 10.3997/2214-4609-pdb.186.INF01.
- Ramakrishna, T.S., 2006. Geophysical practice in mineral exploration and mapping. *Memoirs of the Geological Society of India*.
- Reid, A.B., Allsop, J.M., Granser, H., Millett, A.J. and Somerton, I.W., 1990. Magnetic interpretation in three dimensions using Euler deconvolution. *Geophysics*, vol. 55, pp. 80–91.
- Reid, A.B. and Thurston, J.B., 2014. The structural index in gravity and magnetic interpretation. *Geophysics*, vol. 79, no. 4, pp. J61–J66.
- Reynolds, J.M., 1997a. *An Introduction to Applied and Environmental Geophysics*. New York: John Wiley and Sons, p. 806.
- Reynolds, J.M., 1997b. *An Introduction to Applied and Environmental Geophysics*. Chichester: Wiley, p. 796.
- Reynolds, J.M., 1998. *An Introduction to Applied and Environmental Geophysics*. Chichester, New York, Weinheim, Brisbane, Singapore, and Toronto: Wiley and Sons.
- Roberts, R., Hinze, W. and Leap, D., 1990. Application of the gravity method to the investigation of a landfill in glaciated mid-continent USA: A case study. In *Geotechnical and Environmental Geophysics*, vol. 2, Ward, S.H. Tulsa, OK: Society of Exploration Geophysicists, pp. 253–9.
- Roy, L., 2001. Short note: Source geometry identification by simultaneous use of structural index and shape factor. *Geophysical Prospecting*, vol. 49, pp. 159–164.
- Saibi, H., 2018. Microgravity and Its Applications in Geosciences. *Gravity Geoscience Applications, Industrial Technology and Quantum Aspect*, DOI: 10.5772/intechopen.71223.
- Sharma, P.V., 1997. *Environmental and Engineering Geophysics*. Cambridge and New York, NY: Cambridge University Press, p. 475.
- Sayyab, A. and Valek, R., 1968. Pattern and general properties of the gravity field of Iraq. *23rd Int. Geol. Congo Czechoslovakia*, vol. 5, pp. 129–42.
- Seguin, M.K., Gupta, O.P., Nagrajan, V., Jain, A.K., Sen, N., Mahapatra, S. and Mishra, O.P., 1994. Identification and delineation of abandoned, unknown and unsurveyed galleries in Raniganj coalfield by geophysical methods, Exploration Technology for modern coal mines: Coal exploration. *Seminar*, vol. VI-49–VI-70.

- Seguin, M.K., Sen, N., Mahapatra, S., Mishra, O.P., Gupta, O.P., Nagarajan, V., Jain, A.K. and Akhauri, H.M., 1995. Identification and delineation of abandoned, unknown and unsurveyed underground colliery workings in Raniganj coalfield: Geophysical studies. Final report submitted to the International Development Research Center (IDRC), Canada, Ministry of External Affairs of Canada & Ministry of Science and Technology, Govt. of India, p. 165.
- Silva, J., Teixeira, W. and Barbosa, V., 2008. Gravity data as a tool for landfill study. *Environmental Geology*, DOI: 10.1007/s00254-008-1353-6.
- Strahler, A.N., 1963. *The Earth Sciences*. New York: Harper and Row.
- Sundaram, R.M., Surendranath, M. and Reddy, U.S.N., 2007. Revised Gravity Maps of India (2006 series)—a Cartographic Perspective. Proceeding Map World Forum, January 22–25, 2007, Hyderabad, India.
- Talwani, M., Worzel, J.L. and Landisman, M., 1959. Rapid gravity computations for two- dimensional bodies with applications to the Mendocino submarine fracture zone. *Journal of Geophysical Research*, vol. 64, pp. 49–59.
- Telford, W.M., Geldart, L.P. and Sheriff, R.E., 1990. *Applied Geophysics*. Cambridge: Cambridge University Press.
- Thompson, D.T., 1982. EULDPH: A new technique for making computer-assisted depth estimates from magnetic data. *Geophysics*, vol. 47, pp. 31–37. 38.
- Tiwari, V.M., Ravi Kumar, M. and Mishra, D.C., 2013. Long wavelength gravity anomalies over India: Implications on Lithospheric Structure. *Journal of Asian Earth Sciences*, DOI: 10.1016/j.jseas.2013.03.011.
- Ulrich, T.J., 1968. Effect of wavelength filtering on the shape of the residual anomaly. *Geophysics*, vol. 33, p. 1015.
- Zahorec, P., Mikuska, J., Papco, J. and Pasteka, R., 2023. The comparison of different methods of determining the rock density from gravity data. *ResearchGate, Article: In Contributions to Geophysics and Geodesy*, vol. 53, no. 4, pp. 353–375.

4 Magnetic Method

4.1 INTRODUCTION TO MAGNETIC EXPLORATION: HISTORICAL OVERVIEW

Unlike the gravitational observations described in the previous chapter, man has been systematically observing the Earth's magnetic field for almost 500 years. Sir William Gilbert published the first scientific treatise on the Earth's magnetic field, titled *De magnete*. In this work, Gilbert showed that the reason compass needles point toward the Earth's North Pole is the Earth's magnetism, indicating that the Earth behaves as a giant bar magnet. William Gilbert (1540–1603) recorded terrestrial magnetism using a lodestone or leading stone. Earth's magnetic field is equivalent to a permanent magnetic lying N–S direction near the rotational axis. Geophysical exploration using measurements of the Earth's magnetic field was employed earlier than any other geophysical technique. Von Werde located deposits of ore by mapping variations in the magnetic field in 1843. In 1879, Thalen published the first geophysical manuscript, titled *The Examination of Iron Ore Deposits by Magnetic Measurements*. This method is the fastest and cheapest geophysical technique mostly used to locate iron ore deposits, and that, too, only magnetite. Mineral exploration involves the mapping of geological structures, such as faults, contacts, fractures, shear zones, and acid and basic igneous suits. Common uses of magnetometers include the following (Table 4.1).

4.1.1 SIMILARITIES BETWEEN GRAVITY AND MAGNETICS

Geophysical investigations employing observations of the Earth's magnetic field have much in common with those employing observations of the Earth's gravitational field. Thus, you will find that your previous exposure to, and the intuitive understanding you developed from using, gravity will greatly assist you in understanding the use of magnetics. In particular, some of the most striking similarities between the two methods include:

- Geophysical exploration techniques that employ both gravity and magnetics are passive. By this we simply mean that when using these two methods, we measure a naturally occurring field of the Earth: either the Earth's gravitational or magnetic fields. The gravity and magnetics methods are often referred to as *potential methods*,¹ and the gravitational and magnetic fields we measure are referred to as *potential fields*.

TABLE 4 1

Different Ranges of Applications for Magnetic Surveying

Method	Applications
Magnetic surveying	Locating buried tanks and drums
	Fault studies
	Mineral exploration
	Geothermal exploration
	Mapping buried utilities, pipelines
	Buried foundations, fire pits for archaeological studies

- Identical physical and mathematical representations can be used to understand magnetic and gravitational forces. For example, the fundamental element used to define the gravitational force is the point mass. An equivalent representation is used to define the force derived from the fundamental magnetic element. Instead of being called a *point mass*, however, the fundamental magnetic element is called a *magnetic monopole*. Mathematical representations for the point mass and the magnetic monopole are identical.
- The acquisition, reduction, and interpretation of gravity and magnetic observations are very similar.

4.1.2 DIFFERENCES BETWEEN GRAVITY AND MAGNETICS

Unfortunately, despite these similarities, there are several significant differences between gravity and magnetic exploration. By and large, these differences make the qualitative and quantitative assessment of magnetic anomalies more difficult and less intuitive than gravity anomalies.

- The fundamental parameter that controls gravity variations of interest to us as exploration geophysicists is rock density. The densities of rocks and soils vary little from place to place near the surface of the Earth. The highest densities we typically observe are about 3.0 gm/cc, and the lowest densities are about 1.0 gm/cc. The fundamental parameter controlling the magnetic field variations of interest to us, magnetic susceptibility, on the other hand, can vary as much as four to five orders of magnitude.² This variation is not only present amongst different rock types, but wide variations in susceptibility also occur within a given rock type. Thus, it will be extremely difficult with magnetic prospecting to determine rock types on the basis of estimated susceptibilities.
- Unlike gravitational force, which is always attractive, magnetic force can be either attractive or repulsive. That is, mathematically, monopoles can assume either positive or negative values.
- Unlike the gravitational case, single magnetic point sources (monopoles) can never be found alone in the magnetic case. Rather, monopoles always occur in pairs. A pair of magnetic monopoles, referred to as a *dipole*, always consists of one positive monopole and one negative monopole.
- A properly reduced gravitational field is always generated by subsurface variations in rock density. A properly reduced magnetic field, however, can have as its origin at least two possible sources. It can be produced via an induced magnetization, or it can be produced via a remanent magnetization. For any given set of field observations, both mechanisms probably contribute to the observed field. It is difficult, however, to distinguish between these possible production mechanisms from field observations alone.
- Unlike the gravitational field, which does not change significantly with time,³ the magnetic field is highly time-dependent.

4.2 MAGNETIC MONOPOLES

Recall that the gravitational force exerted between two point masses of mass m_1 and m_2 separated by a distance r is given by Newton's law of gravitation, which is written as

$$F_g = \frac{Gm_1m_2}{r^2}$$

where G is the gravitational constant. This law, in other words, simply states that the gravitational force exerted between two bodies decreases as one over the square of the distance separating the bodies. Since mass, distance, and the gravitational constant are always positive values, the gravitational force is always an attractive force.

Charles-Augustin de Coulomb, in 1785, showed that the force of attraction or repulsion between electrically charged bodies and between magnetic poles also obeys an inverse square law like that derived for gravity by Newton. To make the measurements necessary to prove this, Coulomb (independently of John Michell) invented the torsion balance.

The mathematical expression for the magnetic force experienced between two magnetic monopoles is given by

$$F_m = \frac{1}{\mu} \frac{p_1 p_2}{r^2}$$

where μ is a constant of proportionality known as the ‘magnetic permeability’, p_1 and p_2 are the strengths of the two magnetic monopoles, and r is the distance between the two poles. In form, this expression is identical to the gravitational force expression written earlier. There are, however, two important differences:

- Unlike the gravitational constant, G , the magnetic permeability, μ , is a property of the material in which the two monopoles p_1 and p_2 are located. If they are in a vacuum, μ is called the magnetic permeability of free space.
- Unlike m_1 and m_2 , p_1 and p_2 can be either positive or negative in sign. If p_1 and p_2 have the same sign, the force between the two monopoles is repulsive. If p_1 and p_2 have opposite signs, the force between the two monopoles is attractive.

4.2.1 FORCES ASSOCIATED WITH MAGNETIC MONOPOLES

Given that the magnetic force applied to one magnetic monopole by another magnetic monopole is given by Coulomb’s equation, what does the force look like? Assume that there is a negative magnetic pole, $p_1 < 0.0$, located at a point $x = -1$ and $y = 0$. Now, let us take a positive magnetic pole, $p_2 > 0.0$; move it to some location (x, y) ; and measure the strength and the direction of the magnetic force field. We will plot this force as an arrow in the direction of the force with a length indicating the strength of the force. Repeat this by moving the positive pole to a new location. After doing this at many locations, you will produce a plot similar to the one shown in what follows (Wikipedia; www.sciencefacts.net).

As described by Coulomb’s equation, the size of the arrows should decrease as one over the square of the distance between the two magnetic poles,⁴ and the direction of the force acting on p_2 is always in the direction toward p_1 (the force is attractive).⁵

If instead p_1 is a positive pole located at $x = 1$, the plot of the magnetic force acting on p_2 is the same as that shown earlier, except that the force is always directed away from p_1 (the force is repulsive) (Wikipedia; www.sciencefacts.net).

4.2.2 MAGNETIC DIPOLES

So far, everything seems pretty simple and directly comparable to gravitational forces, albeit with attractive and repulsive forces existing in the magnetic case when only attractive forces existed in the gravitational case. Now things start getting a bit more complicated. The magnetic monopoles that we have been describing have never actually been observed. Rather, the fundamental magnetic element appears to consist of two magnetic monopoles, one positive and one negative, separated by some distance. This fundamental magnetic element consisting of two monopoles is called a *magnetic dipole*.

Now let us see what the force looks like from this fundamental magnetic element, the magnetic dipole. Fortunately, we can derive the magnetic force produced by a dipole by considering the force produced by two magnetic monopoles. In fact, this is why we began our discussion on magnetism by

looking at magnetic monopoles. If a dipole simply consists of two magnetic monopoles, you might expect that the force generated by a dipole is simply the force generated by one monopole added to the force generated by a second monopole. This is exactly right. Grant and Philips (2008) plotted the magnetic forces associated with two magnetic monopoles. These are reproduced in what follows on the same figure as the red and purple arrows (Grant and Philips, 2008).

If we add these forces together using vector addition, we get the green arrows. These green arrows now indicate the force associated with a magnetic dipole consisting of a negative monopole at $x = -1$, labelled S , and a positive monopole at $x = 1$, labelled N . Shown in the following are the force arrows for this same magnetic dipole without the red and purple arrows indicating the monopole forces (Grant and Philips, 2008).

The force associated with this fundamental element of magnetism, the magnetic dipole, now looks more complicated than the simple force associated with gravity. Notice how the arrows describing the magnetic force appear to come out of the monopole labelled N and into the monopole labelled S . You may recognize this force distribution. It is nothing more than the magnetic force distribution observed around a simple bar magnet. In fact, a bar magnet can be thought of as nothing more than two magnetic monopoles separated by the length of the magnet. The magnetic force appears to originate out of the North Pole (N) of the magnet and to terminate at the South Pole (S) of the magnet.

4.2.3 FIELD LINES FOR A MAGNETIC DIPOLE

Another way to visualize the magnetic force field associated with a magnetic dipole is to plot the field lines for the force. Field lines are nothing more than a set of lines drawn such that they are everywhere parallel to the direction of the force you are trying to describe, in this case, the magnetic force. Shown in what follows is the spatial variation of the magnetic force (green arrows)⁶ associated with a magnetic dipole and a set of field lines (red lines) describing the force (Stephen, 2011).

Notice that the red lines representing the field lines are always parallel to the force directions shown by the green arrows. The number and spacing of the red lines that we have chosen to show are arbitrary, except for one factor. The position of the red lines shown has been chosen to qualitatively indicate the relative strength of the magnetic field. Where adjacent red lines are closely spaced, such as near the two monopoles (blue and yellow circles) comprising the dipole, the magnetic force is large. The greater the distance between adjacent red lines, the smaller the magnitude of the magnetic force.

4.2.4 UNITS ASSOCIATED WITH MAGNETIC POLES

The units associated with magnetic poles and the magnetic field are a bit more obscure than those associated with the gravitational field. From Coulomb's expression, we know that force must be given in newtons, N , where a newton is 1 kg-m/s^2 . We also know that distance has the units of meters (m). Permeability μ is defined as a unitless constant. The units of pole strength are defined such that if the force F is 1 N and the two magnetic poles are separated by 1 m, each of the poles has a strength of 1 Amp-m (ampere-meters). In this case, the poles are referred to as *unit poles*.

$$H = \frac{F_m}{p_2} = \frac{p_1}{\mu r^2}$$

The magnetic field strength H is defined as the force per unit pole strength exerted by a magnetic monopole p_1 . H is nothing more than Coulomb's expression divided by p_2 . The magnetic field strength H is the magnetic analogue to the gravitational acceleration g .

Given the units associated with force, N , and magnetic monopoles, *Amp-m*, the units associated with magnetic field strength are newtons per ampere-meter, $N/(\text{Amp-m})$. A $N/(\text{Amp-m})$ is referred to as a tesla (T), named after the renowned inventor Nikola Tesla.

When describing the magnetic field strength of the Earth, it is more common to use units of nanoteslas (nT), where one nanotesla is 1 billionth of a tesla. The average strength of the Earth's magnetic field is about 50,000 nT. A *nanotesla* is also commonly referred to as a *gamma*.

4.3 MAGNETIZATION OF MATERIALS

4.3.1 MAGNETIC INDUCTION ($B = \mu H$)

When a magnetic material, say, iron, is placed within a magnetic field, H , the magnetic material will produce its own magnetization. This phenomenon is called *induced magnetization*. In practice, the induced magnetic field (that is, the one produced by the magnetic material) will look like it is being created by a series of magnetic dipoles located within the magnetic material and oriented parallel to the direction of the inducing field H (Brown, William Fuller, Jr., 1962). The strength of the magnetic field induced by the magnetic material due to the inducing field is called the intensity of magnetization, I , or dipole moment per unit volume.

The true behaviour of induced magnetization may be investigated by placing a sample in a coil. $B = \mu H$ until the sample is *saturated*, after which a further increase in H produces no further increase in B . When H is returned to 0, some magnetization still remains. This is called *remanent magnetization*. This pattern continues and forms a *hysteresis loop*. It shows how a sample can stay magnetized after the magnetizing force is gone (Kearey et al., 2002).

4.3.2 MAGNETIC SUSCEPTIBILITY $\left(k = \frac{I}{H}\right)$

The intensity of magnetization, I , is related to the strength of the inducing magnetic field, H , through a constant of proportionality, k , known as the magnetic susceptibility.

The magnetic susceptibility is a unitless constant that is determined by the physical properties of the magnetic material. It can take on either positive or negative values. Positive values imply that the induced magnetic field, I , is in the same direction as the inducing field, H . Negative values imply that the induced magnetic field is in the opposite direction as the inducing field. In magnetic prospecting, susceptibility is the fundamental material property whose spatial distribution we are attempting to determine. In this sense, magnetic susceptibility is analogous to density in gravity surveying.

4.4 MECHANISMS FOR INDUCED MAGNETIZATION

The nature of material magnetization is, in general, complex, governed by atomic properties. There are three types of magnetic materials: paramagnetic, diamagnetic, and ferromagnetic.

4.4.1 DIAMAGNETISM

Discovered by Michael Faraday in 1846, this form of magnetism is a fundamental property of all materials and is caused by the alignment of magnetic moments associated with orbital electrons⁷ in the presence of an external magnetic field. For those elements with no unpaired electrons in their outer electron shells, this is the only form of magnetism observed. The susceptibilities of diamagnetic materials are relatively small and negative. Quartz and salt are two common diamagnetic Earth materials.

4.4.2 PARAMAGNETISM

This is a form of magnetism associated with elements that have an odd number of electrons in their outer electron shells. Paramagnetism is associated with the alignment of electron spin directions in the presence of an external magnetic field. It can only be observed at relatively low temperatures.

The temperature above which paramagnetism is no longer observed is called the *Curie temperature*, or it is the temperature of demagnetization. The susceptibilities of paramagnetic substances are small and positive. Examples of materials that are paramagnetic are the ${}_{20}\text{Ca}$ – ${}_{28}\text{Ni}$ element series.

4.4.3 FERROMAGNETISM

This is a special case of paramagnetism in which there is an almost-perfect alignment of electron spin directions within large portions of the material referred to as domains. There are only three ferromagnetic elements: iron, cobalt and nickel. Like paramagnetism, ferromagnetism is observed only at temperatures below the Curie temperature. Some examples of Curie temperatures are Fe 750°C, Ni 360°C, and magnetite 578°C. The majority of anomalies in the crust result from magnetite, and so knowledge of the Curie temperature and the geothermal gradient can give information on the depth range of causative bodies.

- *Pure ferromagnetism.* The directions of electron spin alignment within each domain are almost all parallel to the direction of the external inducing field (Somasundaran, 2006). Pure ferromagnetic substances have large (approaching 1) positive susceptibilities. Ferromagnetic minerals do not exist, but iron, cobalt, and nickel are examples of common ferromagnetic elements.
- *Antiferromagnetism.* The directions of electron alignment within adjacent domains are opposite, and the relative abundance of domains with each spin direction is approximately equal (Louis, 1948). The observed magnetic intensity for the material is almost zero. Thus, the susceptibilities of antiferromagnetic materials are almost zero. Hematite is an antiferromagnetic material.
- *Ferrimagnetism.* Like antiferromagnetic materials, adjacent domains produce magnetic intensities in opposite directions. The intensities associated with domains polarized in a direction opposite that of the external field, however, are weaker. The observed magnetic intensity for the entire material is in the direction of the inducing field but is much weaker than that observed for pure ferromagnetic materials (Spaldin, 2010). Thus, the susceptibilities for ferromagnetic materials are small and positive. The most important magnetic minerals are ferromagnetic and include magnetite, titanomagnetite, ilmenite, and pyrrhotite.

4.5 SUSCEPTIBILITIES OF ROCKS AND MINERALS

As we will show later, however, the determination of a material type through knowledge of its susceptibility is an extremely difficult proposition, even more so than by determining a material type through knowledge of its density. Unlike density, notice the large range of susceptibilities not only between varying rocks and minerals but also within rocks of the same type. It is not uncommon to see variations in susceptibility of several orders of magnitude for different igneous rock samples. In addition, like density, there is considerable overlap in the measured susceptibilities. Hence, knowledge of susceptibility alone will not be sufficient to determine rock type, and alternately, knowledge of rock type is often not sufficient to estimate the expected susceptibility. This wide range in susceptibilities implies that spatial variations in the observed magnetic field may be readily related to geological structure. Because variations within any given rock type are also large, however, it will be difficult to construct corrections to our observed magnetic field on assumed susceptibilities, as was done in constructing some of the fundamental gravitational corrections (Bouguer slab correction and topographic corrections). The susceptibilities of various rocks and minerals are shown in the following (Table 4.2) (Grant and West, 1965).

Sedimentary rocks usually have very low k , and sedimentary structures very rarely give large magnetic anomalies. If a large magnetic anomaly occurs in a sedimentary environment, it is usually due to an igneous body at depth. Common causes of magnetic anomalies in dykes, folded or faulted sills, lava flows, basic intrusions, metamorphic basement rocks, and ore bodies that contain

TABLE 4.2
Susceptibilities of Rocks and Minerals

Material	Susceptibility $\times 10^{-3}(\text{SI})^a$
Air	~ 0
Quartz	-0.01
Rock salt	-0.01
Calcite	-0.001 to 0.01
Sphalerite	0.4
Pyrite	0.05 – 5
Hematite	0.5 – 35
Ilmenite	300 – $3,500$
Magnetite	$1,200$ – $19,200$
Limestones	0 – 3
Sandstones	0 – 20
Shales	0.01 – 15
Schist	0.3 – 3
Gneiss	0.1 – 25
Slate	0 – 35
Granite	0 – 50
Gabbro	1 – 90
Basalt	0.2 – 175
Peridotite	90 – 200

magnetite all generate large-amplitude magnetic anomalies. Other targets suitable for study using magnetics are disturbed soils at shallow depth, fire pits, and kilns, all of which are of interest in archaeological studies. Magnetic anomalies can be used to get the depth of basement rocks and, hence, sedimentary thickness.

4.6 REMANENT MAGNETIZATION

If the magnetic material has relatively large susceptibilities, or if the inducing field is strong, the magnetic material will retain a portion of its induced magnetization even after the induced field disappears. This remaining magnetization is called *remanent magnetization*. Remanent magnetization is the component of the material's magnetization that solid Earth geophysicists use to map the motion of continents and ocean basins resulting from plate tectonics. Rocks can acquire a remanent magnetization through a variety of processes:

Chemical remanent magnetization (CRM). This is acquired as a result of chemical grain accretion or alteration and affects sedimentary and metamorphic rocks.

Detrital remanent magnetization (DRM). This is acquired as particles settle in the presence of the Earth's field. The particles tend to orient themselves as they settle.

Isothermal remanent magnetism (IRM). This is the residual magnetic field left when an external field is applied and removed, for example, lightning.

Thermoremanent magnetization (TRM). This is acquired when rock cools through the Curie temperature, and characterizes most igneous rocks. It is the most important kind of magnetization for palaeomagnetic dating.

Viscous remanent magnetization (VRM). Rocks acquire this after long exposure to an external magnetic field, and it may be important in fine-grained rocks.

A simple example, however, will illustrate the concept. As a volcanic rock cools, its temperature decreases past the Curie temperature. At the Curie temperature, the rock, being magnetic, begins to produce an induced magnetic field. In this case, the inducing field is the Earth's magnetic field. As the Earth's magnetic field changes with time, a portion of the induced field in the rock does not change but remains fixed in a direction and strength reflective of the Earth's magnetic field at the time the rock cools through its Curie temperature. This is the remanent magnetization of the rock—the recorded magnetic field of the Earth at the time the rock cooled past its Curie temperature.

The only way you can measure the remanent magnetic component of a rock is to take a sample of the rock back to the laboratory for analysis. This is time-consuming and expensive. As a result, in exploration geophysics, we typically assume there is no remnant magnetic component in the observed magnetic field. Clearly, however, this assumption is wrong and could possibly bias our interpretations. The remanent magnetization will be measured using an astatic or spinner magnetometer, which measures the magnetism of samples in the absence of the Earth's field.

4.7 MAGNETIC FIELD NOMENCLATURE

As you can see, although we started by comparing the magnetic field to the gravitational field, the specifics of magnetism are far more complex than that of gravitation. Despite this, it is still useful to start from the intuition you have gained through your study of gravitation when trying to understand magnetism. Before continuing, however, we need to define some of the relevant terms we will use to describe the Earth's magnetic field.

When discussing gravity, we really did not talk much about how we describe gravitational acceleration. To some extent, this is because such a description is almost obvious; gravitational acceleration has some size (measured in geophysics with a gravimeter in mGal), and it is always acting downward (in fact, it is how we define *down*). Because the magnetic field does not act along any such easily definable direction, Earth scientists have developed a nomenclature to describe the magnetic field at any point on the Earth's surface.

At any point on the Earth's surface, the magnetic field, F^* , has some strength and points in some direction (Figure 4.1; Nettleton, 1942). The following terms are used to describe the direction of the magnetic field.

- *Declination (D)*. The angle between north and the horizontal projection of F . This value is measured positive through east and varies from 0 to 360 degrees.
- *Inclination (I)*. The angle between the surface of the Earth and F . Positive inclinations indicate F is pointed downward, while negative inclinations indicate F is pointed upward. Inclination varies from -90 to 90 degrees.
- *Magnetic equator*. The location around the surface of the Earth where the Earth's magnetic field has an inclination of zero (the magnetic field vector F is horizontal). This location does not correspond to the Earth's rotational equator.
- *Magnetic poles*. The locations on the surface of the Earth where the Earth's magnetic field has an inclination of either plus or minus 90 degrees (the magnetic field vector F is vertical). These locations do not correspond to the Earth's North and South Poles.

* In this context, and throughout the remainder of this chapter, 'F' includes contributions from the Earth's main magnetic field (the main magnetic field refers to that portion of the Earth's magnetic field that is believed to be generated within the Earth's core. It constitutes the largest portion of the magnetic field and is the field that acts to induce magnetization in crustal rocks that we are interested in for exploration applications, induced magnetization from crustal sources, and any contributions from sources external to the Earth (the inducing field).

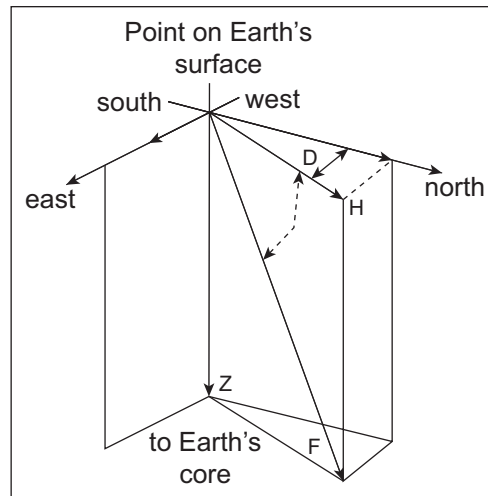


FIGURE 4.1 Magnetic elements.

In addition to F , I , and D , there are Z , the vertical component reckoned positive downward; H , the horizontal component always positive; and X and Y , components of H to the north and east, respectively. From the diagram:

$$F^2 = H^2 + Z^2 = X^2 + Y^2 + Z^2$$

$$Z = F \sin I$$

where $\sin I = \frac{Z}{F}$, $X = H \cos D$, and $Y = H \sin D$. From this relation we have $\tan D = \frac{Y}{X}$.

4.8 EARTH'S MAGNETIC FIELD

Around 90% of the Earth's magnetic field looks like a magnetic field that would be generated from a dipolar magnetic source located at the centre of the Earth and aligned with the Earth's rotational axis. This first-order description of the Earth's magnetic field was first given by Sir William Gilbert in 1600. The strength of the magnetic field at the poles is about $50,000 \text{ nT}$. If this dipolar description of the field were complete, then the magnetic equator would correspond to the Earth's equator and the magnetic poles would correspond to the geographic poles (Figure 4.2).

The remaining 10% of the magnetic field cannot be explained in terms of simple dipolar sources. Complex models of the Earth's magnetic field have been developed and are available. Shown in the following is a sample of one of these models generated by the USGS (National Geophysical Data Center at NOAA; US/UK World Magnetic Model; Main Field Total Intensity [F]). If the Earth's field were simply dipolar, with the axis of the dipole oriented along the Earth's rotational axis, all declinations would be '0' degrees (the field would always point towards the north). As can be seen, the observed declinations are quite complex.

As observed on the surface of the Earth, the magnetic field can be broken into three separate components.

- *Main field.* This is the largest component of the magnetic field and is believed to be caused by electrical currents in the Earth's fluid outer core. For exploration work, this field acts as the inducing magnetic field.

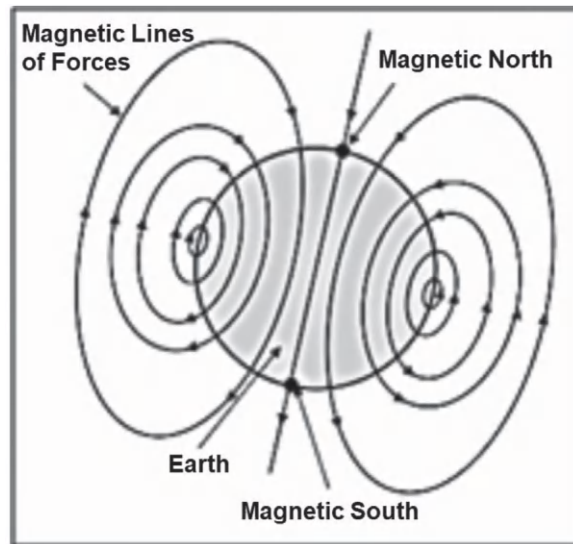


FIGURE 4.2 Geomagnetic field of the Earth.

Source: Wikipedia.

- *External magnetic field.* This is a relatively small portion of the observed magnetic field that is generated from magnetic sources external to the Earth. This field is believed to be produced by the interactions of the Earth's ionosphere with the solar wind. Hence, some temporal variations associated with the external magnetic field are correlated to solar activity.
- *Crustal field.* This is the portion of the magnetic field associated with the magnetism of crustal rocks. This portion of the field contains magnetism caused both by the induction from the Earth's main magnetic field and by remanent magnetization.

The figure shown earlier (National Geophysical Data Center at NOAA; US/UK World Magnetic Model; Main Field Total Intensity [F]) was constructed to emphasize the characteristics of the main magnetic field. Although this portion of the field is in itself complex, it is understood quite well. Models of the main field are available and can be used for data reduction.

4.9 MAGNETICS AND GEOLOGY: A SIMPLE EXAMPLE

This is all beginning to get a bit complicated. What are we actually going to observe, and how is this related to geology? The portion of the magnetic field that we have described as the main magnetic field is believed to be generated in the Earth's core. There are a variety of reasons that geophysicists believe that the main field is being generated in the Earth's core, but these are not important for our discussion. In addition to these core sources of magnetism, rocks exist near the Earth's surface that are below their Curie temperature and, as such, can exhibit induced as well as remnant magnetization.⁹

Therefore, if we were to measure the magnetic field along the surface of the Earth, we would record magnetization due to both the main and induced fields. The induced field is the one of interest to us because it relates to the proximity of our measurement to rocks' high or low magnetic susceptibility. If our measurements are taken near rocks of high magnetic susceptibility, we will, in general,¹⁰ record magnetic field strengths that are larger than if our measurements were taken at a great distance from rocks of high magnetic susceptibility. Hence, like gravity, we can potentially locate subsurface rocks having high magnetic susceptibilities by mapping variations in the strength of the magnetic field at the Earth's surface.

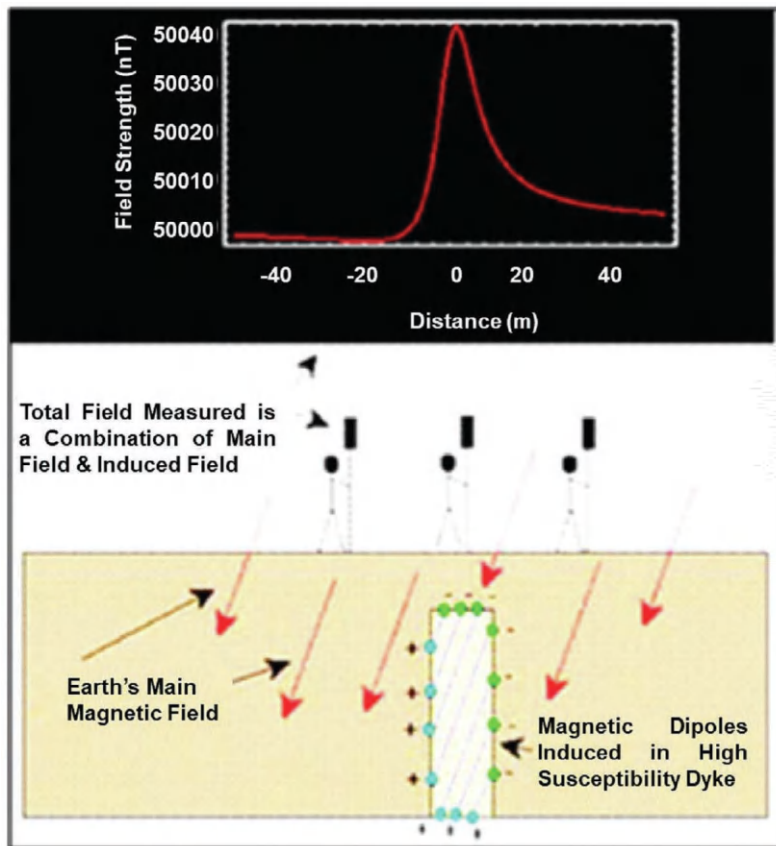


FIGURE 4.3 Magnetic effects over dyke trending north-east.

Suppose we have a buried dyke with a susceptibility of 0.001 surrounded by sedimentary rocks with no magnetic susceptibility (Figure 4.3; Kamar, 2013). The dyke in this example is 3 m wide, is buried 5 m deep, and trends to the north-east. To find the dyke, we could measure the strength of the magnetic field (in this case, along an east–west trending line). As we approach the dyke, we would begin to observe the induced magnetic field associated with the dyke in addition to the Earth’s main field.

Thus, we could determine the location of the dyke, and possibly its dimensions, by measuring the spatial variation in the strength of the magnetic field. There are several things to notice about the magnetic anomaly produced by this dyke:

- Like a gravitational anomaly associated with a high-density body, the magnetic anomaly associated with the dyke is localized to the region near the dyke. The size of the anomaly rapidly decays with distance away from the dyke.
- Unlike the gravity anomaly we would expect from a higher-density dyke, the magnetic anomaly is not symmetric about the dyke’s midpoint, which is at a distance of zero in the earlier example. Not only is the anomaly shaped differently to the left and to the right of the dyke, but the maximum anomaly is also not centred at the centre of the dyke. These observations are, in general, true for all magnetic anomalies. The specifics of this generalization, however, will depend on the shape and orientation of the magnetized body, its location (bodies of the same shape and size will produce different anomalies when located at different places), and the direction in which the profile is taken.

- The size of the anomaly produced by this example is about 40 nT. This is a pretty good-sized anomaly. It is not uncommon to look for anomalies as small as a few nT. Thus, we must develop surveying techniques to reduce systematic and random errors to smaller than a few nT.

4.10 TEMPORAL VARIATIONS OF THE EARTH'S MAGNETIC FIELD: OVERVIEW

Like the gravitational field, the magnetic field varies with time. When describing temporal variations of the magnetic field, it is useful to classify these variations into one of three types, depending on their rate of occurrence and source. Please note explicitly that the temporal variations in the magnetic field that we will be discussing are those that have been observed directly in human history. However, the best-known temporal variation, magnetic polarity reversals, while important in the study of Earth history, will not be considered in this discussion. We will, however, consider the following three temporal variations:

- *Secular variations.* These are long-term (changes in the field that occur over years) variations in the main magnetic field that are presumably caused by fluid motion in the Earth's outer core. Because these variations occur slowly with respect to the time of completion of a typical exploration magnetic survey, these variations will not complicate data reduction efforts.
- *Diurnal variations.* These are variations in the magnetic field that occur over the course of a day and are related to variations in the Earth's external magnetic field. This variation can be in the order of 20 to 30 nT per day and should be accounted for when conducting exploration magnetic surveys.
- *Magnetic storms.* Occasionally, magnetic activity in the ionosphere will abruptly increase. The occurrence of such storms correlates with enhanced sunspot activity. The magnetic field observed during such time is highly irregular and unpredictable, having amplitude changes as large as 1,000 nT. Exploration magnetic surveys should not be conducted during magnetic storms.

4.10.1 SECULAR VARIATIONS OF THE EARTH'S MAGNETIC FIELD

The fact that the Earth's magnetic field varies with time was well established several centuries ago. In fact, this is the primary reason that permanent magnetic observatories were established, from which we have learned how the magnetic field has changed over the past few centuries. Many sources of historical information are available. Shown in what follows is a plot of the declination and inclination of the magnetic field around Britain from the years 1500 through 1900 (Jackson et al., 2000).

At this one location, you can see that, over the past 400 years, the declination has varied by almost 37° , while the inclination has varied by as much as 13° . These changes are generally assumed to be associated with the Earth's main magnetic field. That is, these are changes associated with that portion of the magnetic field believed to be generated in the Earth's core. Solid Earth geophysicists are very interested in studying these secular variations because they can be used to understand the dynamics of the Earth's core.

To understand these temporal variations and to quantify the rate of variability over time, standard reference models are constructed from magnetic observatory observations about every five years. One commonly used set of reference models is known as the *International Geomagnetic Reference Field* (IGRF). Based on these models, it is possible to predict the portion of the observed magnetic field associated with the Earth's main magnetic field at any point on the Earth's surface, both now and for several decades in the past. Because the main magnetic field as described by these secular variations changes slowly with respect to the time it takes us to complete our exploration magnetic survey, this type of temporal variation is of little importance to us.

4.10.2 DIURNAL VARIATIONS OF THE EARTH'S MAGNETIC FIELD

Of more importance to exploration geophysical surveys are the daily, or diurnal, variations of the Earth's magnetic field. These variations were first discovered in 1722 in England, when it was also noted that these daily variations were larger in summer than in winter. The following plot shows typical variations in the magnetic data recorded at a single location (Boulder, Colorado) over a time period of two days (Geomagnetic Indices Bulletin, 1990). Although there are high-frequency components to this variation, notice that the dominant trend is a slowly varying component with a period of about 24 hours. In this location, at this time, the amplitude of this daily variation is about 20 nT.

These variations are believed to be caused by electric currents induced in the Earth from an external source. In this case, the external source is believed to be electric currents in the upper atmosphere, or the ionosphere. These electric currents in the ionosphere are, in turn, driven by solar activity. Given the size of these variations, the size of the magnetic anomalies we would expect in a typical geophysical survey, and the fact that surveys could take several days or weeks to complete, it is clear that we must account for diurnal variations when interpreting our magnetic data.

4.10.3 MAGNETIC STORMS

In addition to the relatively predictable and smoothly varying diurnal variations, there can be transient, large-amplitude (up to 1,000 nT) variations in the field that are referred to as *magnetic storms* (Geomagnetic Indices Bulletin, 1990).

The frequency of these storms correlates with sunspot activity. Based on this, some prediction of magnetic storm activity is possible. The most intense storms can be observed globally and may last for several days. The figure (Geomagnetic Indices Bulletin, 1990) shows the magnetic field recorded at a single location during such a transient event. Although the magnetic storm associated with this event is not particularly long-lived, notice that the size of the magnetic field during this event varies by almost 100 nT in a time period shorter than 10 minutes. Exploration magnetic surveys should not be conducted during times of magnetic storms. This is simply because the variations in the field that they can produce are large, rapid, and spatially varying. Therefore, it is difficult to correct them in acquired data.

4.11 MEASURING THE EARTH'S MAGNETIC FIELD

Instruments for measuring aspects of the Earth's magnetic field are among some of the oldest scientific instruments in existence. Magnetic instruments can be classified into two types.

4.11.1 MECHANICAL INSTRUMENTS

These are instruments that are mechanical in nature and usually measure the attitude (its direction or a component of its direction) of the magnetic field. The most common example of this type of instrument is the simple compass. The compass consists of nothing more than a small test magnet that is free to rotate in the horizontal plane (Figure 4.4; Kamar, 2013). Because the positive pole of the test magnet is attracted to the Earth's negative magnetic pole and the negative pole of the test magnet is attracted to the Earth's positive magnetic pole, the test magnet will align itself along the horizontal direction of the Earth's magnetic field. Thus, it provides measurements of the declination of the magnetic field. The earliest known compass was invented by the Chinese at least 2,000 years ago.

Although compasses are the most common type of mechanical device used to measure the horizontal attitude of the magnetic field, other devices have been devised to measure other components of the magnetic field. Most common among these are the dip needle and the torsion magnetometer. The *dip needle*, as its name implies, is used to measure the inclination of the magnetic field. The dip needle may have been the first device used specifically for geophysical exploration, for magnetic ore. The *torsion magnetometer* is a device that can measure, through mechanical means, the strength of the vertical (or horizontal) component of the magnetic field.

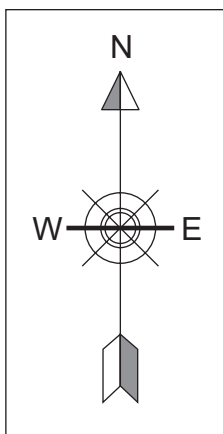


FIGURE 4.4 Magnetic mechanical instrument.

4.11.2 MAGNETOMETERS

Magnetometers are instruments, usually operating non-mechanically, that are capable of measuring the amplitude (strength) of the magnetic field, or of a component of the field. The first advances in designing these instruments were made during World War II, when fluxgate magnetometers were developed for use in submarine detection from low-flying aircraft. Since that time, several other magnetometer designs have been developed that include the proton precession and alkali-vapour magnetometers.

4.11.2.1 Fluxgate Magnetometer

Today it is used for making borehole measurements, and the transducers are found in electronic compasses and in laboratory devices for measuring remnant magnetization. Much airborne magnetic surveying was carried out using fluxgate detectors between 1945 and 1985, and handheld portable devices were used for making vertical-component ground measurements between 1965 and 1985. A schematic of the fluxgate magnetometer is shown in what follows (Nettleton, 1942). Fluxgate magnetometers are capable of measuring the strength of the magnetic field to about 0.5 to 1.0 nT. These are relatively simple instruments to construct; hence, they are relatively inexpensive. Unlike the commonly used gravimeters, fluxgate magnetometers show no appreciable instrument drift with time.

The fluxgate magnetometer is based on what is referred to as the magnetic saturation circuit. Two parallel bars of a ferromagnetic material are placed closely together. The susceptibility of the two bars is large enough so that even the Earth's relatively weak magnetic field can produce near magnetic saturation¹¹ in the bars. Each bar is wound with a primary coil, but the direction in which the coil is wrapped around the bars is reversed. An alternating current (AC) is passed through the primary coils, causing a large, artificial, and varying magnetic field in each coil. This produces induced magnetic fields in the two cores that have the same strengths but opposite orientations at any given time during the current cycle. If the cores are in an external magnetic field, one component of the external field will be parallel to the core axes. As the current in the primary coil increases, the magnetic field in one core will be parallel to the external field and so reinforced by it. The other will be in opposition to the external field and so smaller. The field will reach saturation in one core at a time different from the other core (and fall below saturation, as the current decreases, at a different time).

This difference is sufficient to induce a measurable voltage in a secondary coil that is proportional to the strength of the magnetic field in the direction of the cores. The secondary coil surrounds the two ferromagnetic cores and the primary coil. The magnetic fields induced in the cores by the primary coil produce a voltage potential in the secondary coil. In the absence of an external

field (i.e. if the Earth had no magnetic field), the voltage detected in the secondary coil would be zero because the magnetic fields generated in the two cores have the same strength but are in opposite directions (their effects on the secondary coil exactly cancel). In the presence of an external field component, the behaviour in the two cores differs by an amount which depends on the external field. Thus, the fluxgate magnetometer is capable of measuring the strength of any component of the Earth's magnetic field by simply reorienting the instrument so that the cores are parallel to the desired component. Data acquisition by the fluxgate magnetometer is shown in what follows (Naskar et al., 2018).

4.11.2.2 Proton Precession Magnetometer

For land-based magnetic surveys, the most commonly used magnetometer is the proton precession magnetometer. Unlike the fluxgate magnetometer, the proton precession magnetometer only measures the total amplitude (size) of the Earth's magnetic field. These types of measurements are usually referred to as total field measurements. A schematic of the proton precession magnetometer is shown in the following (Kamar, 2013).

The sensor component of the proton precession magnetometer is a cylindrical container filled with a liquid rich in hydrogen atoms surrounded by a coil. Commonly used liquids include water, kerosene, and alcohol. The sensor is connected by a cable to a small unit in which housed is a power supply, an electronic switch, an amplifier, and a frequency counter. When the switch is closed, a DC current delivered by a battery is directed through the coil, producing a relatively strong magnetic field in the fluid-filled cylinder. The hydrogen nuclei (protons), which behave like minute spinning dipole magnets, become aligned along the direction of the applied field (i.e. along the axis of the cylinder). Power is then cut to the coil by opening the switch. Because the Earth's magnetic field generates a torque on the aligned, spinning hydrogen nuclei, they begin to precess¹² around the direction of the Earth's total field. This precession produces a time-varying magnetic field which induces a small alternating current in the coil. The frequency of the AC current is equal to the frequency of precession of the nuclei. Because the frequency of precession is proportional to the strength of the total field, and because the constant of proportionality is well known, the total field strength can be determined quite accurately.

Like the fluxgate magnetometer, the proton precession magnetometer is relatively easy to construct. Thus, it is also relatively inexpensive. The strength of the total field can be measured down to about 0.1 nT. Like fluxgate magnetometers, proton precession magnetometers show no appreciable instrument drift with time. One of the important advantages of the proton precession magnetometer is its ease of use and reliability. Sensor orientation need only be set to a high angle with respect to the Earth's magnetic field. No precise levelling or orientation is needed. If, however, the magnetic field changes rapidly from place to place, different portions of the cylindrical sensor will be influenced by magnetic fields of various magnitudes, and readings will be seriously degraded. This may occur if the sensor is placed close to a magnetically susceptible material, for instance. Finally, because the signal generated by precession is small, this instrument cannot be used near AC power sources.

4.11.2.2.1 Total Field Measurements

Given the ease of use of the proton precession magnetometer, most exploration geophysical surveys employ this instrument and thus measure only the magnitude of the total magnetic field as a function of position. Surveys conducted using the proton precession magnetometer do not have the ability to determine the direction of the total field as a function of location.

Ignoring for the moment the temporally varying contribution to the recorded magnetic field caused by the external magnetic field, the magnetic field we record with our proton precession magnetometer has two components:

- The *main magnetic field*, or that part of the Earth's magnetic field generated by deep (outer core) sources. The direction and size of this component of the magnetic field at some point on the Earth's surface are represented by the vector labelled F_e in Figure 4.5 (Kamar, 2013).

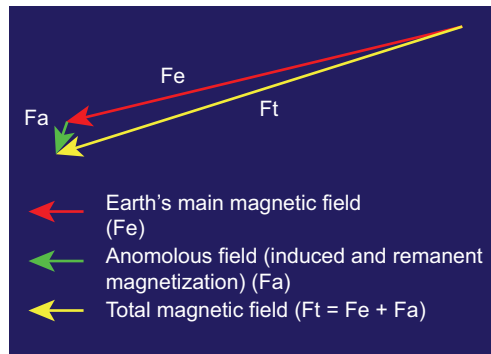


FIGURE 4.5 Direction and component of the Earth's magnetic field.

- The *anomalous magnetic field*, or that part of the Earth's magnetic field caused by magnetic induction of crustal rocks or remanent magnetization of crustal rocks. The direction and size of this component of the magnetic field are represented by the vector labelled F_a in Figure 4.5. The total magnetic field we record, labelled F_t in Figure 4.5, is nothing more than the sum of F_e and F_a . Typically, F_e is much larger than F_a , as is shown in the figure (50,000 nT versus 100 nT). If F_e is much larger than F_a , then F_t will point almost in the same direction as F_e , regardless of the direction of F_a . That is because the anomalous field, F_a , is so much smaller than the main field, F_e , that the total field, F_t , will be almost parallel to the main field.

4.12 FIELD PROCEDURES

4.12.1 MODES OF ACQUIRING MAGNETIC OBSERVATIONS

Magnetic observations are routinely collected using any one of the three different field operational strategies.

4.12.1.1 Airborne

Both fluxgate and proton precession magnetometers can be mounted within or towed behind aircraft, including helicopters. These so-called aeromagnetic surveys are rapid and cost-effective. When relatively large areas are involved, the cost of acquiring 1 km of data from an aeromagnetic survey is about 40% less than the cost of acquiring the same data on the ground. In addition, data can be obtained from areas that are otherwise inaccessible. Among the most difficult problems associated with aeromagnetic surveys is fixing the position of the aircraft at any time. With the development of real-time, differential GPS systems, however, this difficulty has been overcome.

4.12.1.2 Shipborne

Magnetic surveys can also be completed over water by towing a magnetometer behind a ship. Obviously, marine magnetic surveying is slower than airborne surveying. When other geophysical methods are being conducted by ship, however, it may make sense to acquire magnetic data simultaneously.

4.12.1.3 Ground-Based

Like gravity surveys, magnetic surveys are also commonly conducted on foot or with a vehicle. Ground-based surveys may be necessary when the target of interest requires more closely spaced readings than are possible to acquire from the air. In the next discussion, we will concentrate on ground-based surveys. All this discussion, however, could be applied to air- and shipborne surveys also.

Because magnetic surveying is generally far cheaper than other geophysical methods, magnetic observations are commonly used for reconnaissance. These surveys can cover large areas and are used to identify the locations of targets for more detailed investigations. Because of their cost-effectiveness, magnetic surveys usually consist of areal distributions of data instead of single lines of data. We will refer to the collection of geophysical observations over a geographic area as two-dimensional surveys. Data that is collected along a single line of observations will be referred to as one-dimensional surveys.

4.12.2 MAGNETIC CLEANLINESS AND INTERFERENCE

When making total field measurements from which estimates of the subsurface distribution of magnetic susceptibility or the presence of subsurface magnetized bodies are made, it is imperative that factors affecting the recorded field other than these be eliminated or isolated so that they can be removed. We have already discussed several of these added complications, including spatial variations of the Earth's main magnetic field and temporal variations mostly associated with the external magnetic field. In addition to these factors which we cannot control, there are other sources of noise that we can control.

Because any ferromagnetic substance can produce an induced magnetic field in the presence of the Earth's main field, and because modern magnetometers are very sensitive (0.1 nT), the field crew running the magnetic survey must divest itself of all ferrous objects. This includes, but is not limited to, belt buckles, knives, wire-rimmed glasses, etc. In addition, proton precession magnetometers are typically placed on 2 to 3 m poles to remove them from potential noise sources at ground level (GSM-19T, GEM make, Canada, Majumdar et al., 2016, GSI FSP 2015–16, Eastern Region, Kolkata). In addition to noise sources carried by the operators, many sources of magnetic noise may be found in the environment. These can include any ferrous objects, such as houses, fences, railroad rails, cars, rebar in concrete foundations, etc. Finally, when using a proton precession magnetometer, reliable readings will be difficult to obtain near sources of AC power, such as utility lines and transformers. Good observational practice will see visible, non-geological magnetic sources recorded as part of the field observations.

4.12.3 STRATEGIES FOR DEALING WITH TEMPORAL (DRIFT AND DIURNAL) VARIATIONS

Like our gravity observations, magnetic readings taken at the same location at different times will not yield the same results. There are temporal variations in both the Earth's magnetic and gravitational fields.

In acquiring gravity observations, we accounted for this temporal variability by periodically reoccupying a base station and using the variations in this reading to account for instrument drift and temporal variations of the field. We could use the same strategy in acquiring magnetic observations, but this is not routinely done, for the following reasons:

- *Field variations can be more erratic.* Unlike the gravitational field, the magnetic field can vary quite erratically with time, as shown in the following figures (Alfred Wolf, 1940). What this means is that to adequately approximate the temporal variation in the magnetic field by linearly interpolating between base station reoccupations, a very short reoccupation time interval may be required. The shorter the reoccupation interval, the more time is spent at the base station and the longer the survey will take to complete.
- *Instrument drift.* Unlike gravimeters, magnetometers show no appreciable instrument drift.

With these points in mind, most investigators conduct magnetic surveys using two magnetometers. One is used to monitor temporal variations of the magnetic field continuously at a chosen base station, and the other is used to collect observations related to the survey proper. By recording the times at which each magnetic station readings are made and subtracting the magnetic field strength at the base station recorded at that same time, temporal variations in the magnetic field can be eliminated. The resulting field then represents relative values of the variation in total field strength with

respect to the magnetic base station. (For a large-area survey or an airborne survey, the magnetic field may be monitored at several locations, continuously, to protect against spatial variations in the time-varying field.)

4.12.4 SPATIALLY VARYING CORRECTIONS

When reducing gravity observations, there were a host of spatially varying corrections that were applied to the data. These included latitude corrections, elevation corrections, slab corrections, and topography corrections. In principle, all these corrections could be applied to magnetic observations also. In practice, the only corrections routinely made for are spatial variations in the Earth's main magnetic field, which would be equivalent to latitude corrections applied to gravity observations. Why are the other corrections not applied?

Variations in total field strength as a function of elevation are less than 0.015 nT per meter. This variation is generally considered small enough to ignore. Variations in total field strength caused by excess magnetic material in topography could, on the other hand, be quite significant. The problem is the large variation in susceptibilities associated with Earth materials even when those materials are of the same rock type.

Recall that in applying the slab and elevation corrections to our gravitational observations, we had to assume an average density for the rocks making up the corrections. Rock densities do not vary much from rock type to rock type. Density variations of 0.5 gm/cc are large. Variations among different samples of the same rock type vary by even less. Therefore, we can assume an average density for the correction and feel fairly confident that our assumption is reasonable.

Magnetic susceptibilities vary by orders of magnitude even among samples of the same rock type. So how can we choose an average susceptibility on which to base our correction? The answer is, we cannot. Therefore, instead of applying a set of corrections that we know will be wrong, we apply no correction at all to attempt to account for excess material and topography. (For the magnetic equivalent of the Bouguer correction, the solution is even simpler; the magnetic effect of an inductively magnetized, parallel-sided slab is everywhere zero!) The sources of magnetic anomalies could not be very deep, since temperatures below ~40 km should be above the Curie point, the temperature at which rocks lose their magnetic properties ($\approx 550^\circ\text{C}$). Thus, local anomalies must be associated with features in the upper crust.

4.12.5 CORRECTING FOR MAIN FIELD VARIATIONS

Corrections for spatial variations in the strength of the Earth's main magnetic field are referred to as geomagnetic corrections. One commonly used method of accounting for these variations is to use one of the many models of the Earth's main magnetic field that are available. One such set of commonly used models of the main field is referred to as the International Geomagnetic Reference Field (IGRF).

The IGRF models are regularly updated to account for secular variations. Given the latitude and longitude of some point on the Earth's surface, the total field strength of the Earth's main magnetic field can be calculated.

Consider a small two-dimensional survey. A plan view of such a survey is shown in Kamar (2013, p. 29). One commonly used method of applying the main field correction is to linearly interpolate the computed values of the main field at the corners of the survey throughout the survey region. These interpolated values can then be subtracted from the field observations. After applying this correction, you are left with that portion of the magnetic field that cannot be attributed to the Earth's main magnetic field.

This two-dimensional application of linear interpolation is only slightly more complex than the one-dimensional linear interpolation used to reduce our gravity observations. Values of the Earth's main magnetic field are first determined from the IGRF for each corner point of the survey (c_1, c_2, c_3, c_4). To determine the strength of the main field at point p , we first perform two linear interpolations up the edges of the survey in the y direction to determine the values of the field at points t_1 and t_2 . That

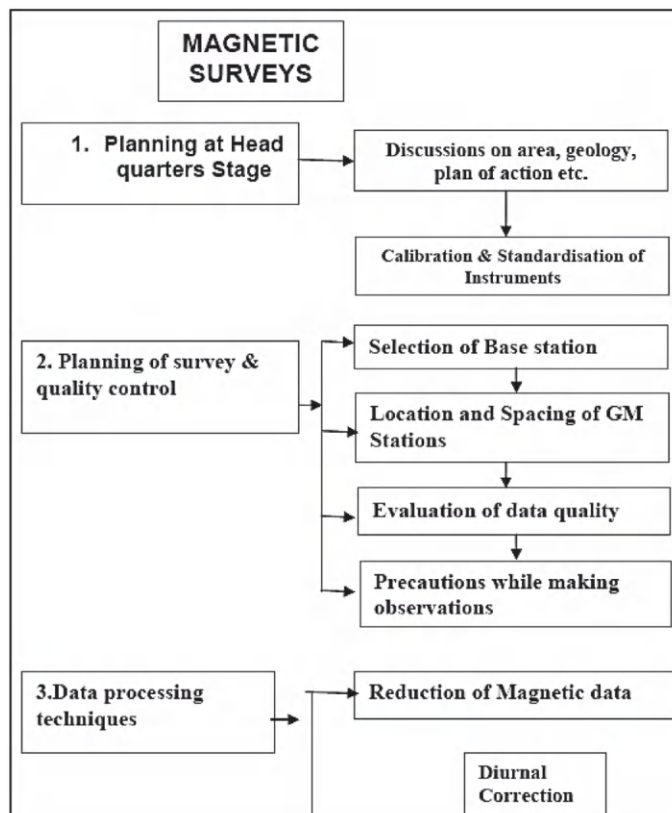
TABLE 4.3**Data Reduction Sheet for Total Field Magnetic Data**

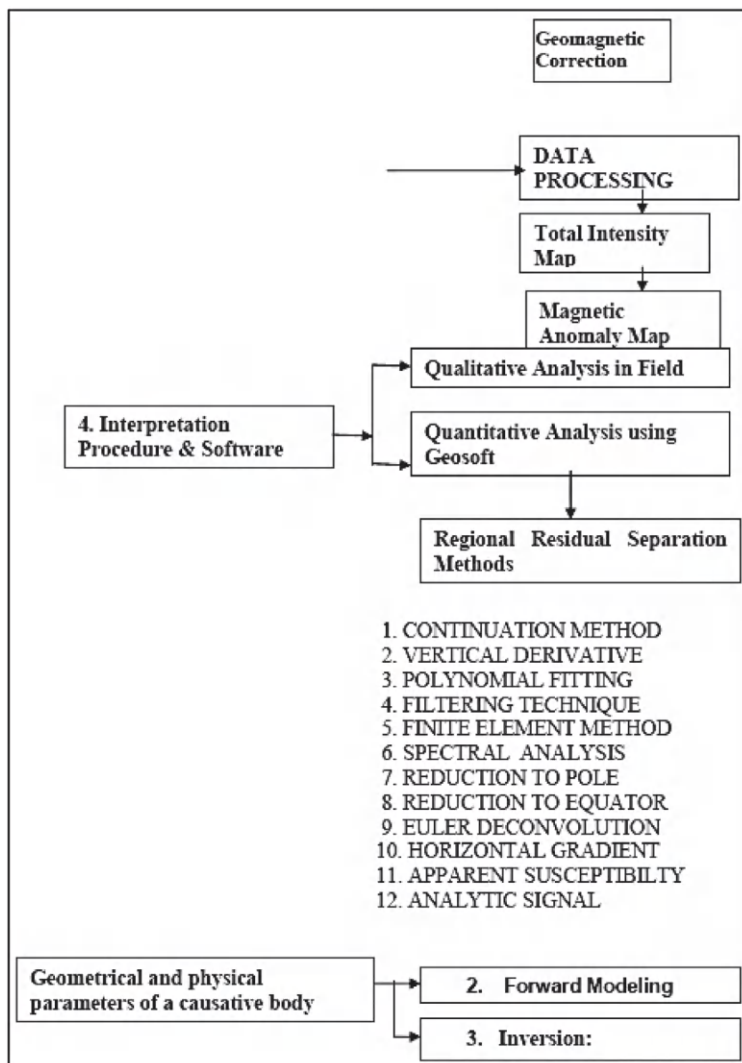
Stn. Id. 1	Long. (Deg) 2	Lat. (Deg) 3	Elev. (m) 4	Diurnal Corrected Field (nT) 5	IGRF Normal Field Epoch . . . (nT) 6	IGRF Inclination	IGRF Declination	Magnetic Anomaly Col. (5) Minus Col. (6) nT
Base								

is, first determine the value of the Earth's main magnetic field at the point t_1 by linearly interpolating between the points c_1 and c_4 . Then determine the value of the main field at the point t_2 by linearly interpolating between the points c_2 and c_3 . Now, linearly interpolate in the x direction between t_1 and t_2 . The result is the two-dimensionally interpolated value of the field at the point p^{13} . Following is the data reduction sheet for total field magnetic data in the magnetic geophysical survey (Table 4.3).

4.13 FORMAT FOR MAGNETIC SURVEYING

The format for magnetic data processing has been suggested, which is essential to maintain the uniformity and quality of the output of magnetic surveying.





4.14 DATA PROCESSING

All the processing techniques, including data quality checks, analysis, management, and archival for gravity data, are also applicable to magnetic data. In addition to those, in the case of magnetic data, reduced-to-pole (RTP) and reduced-to-equator (RTE) techniques may also be applied. The understanding of the density and susceptibility values of the rocks/minerals in the area of study is essential to obtain a meaningful interpretation of the maps and for quantitative interpretations. Parameters like magnetic inclination, dip, declination, and the ambient field (normal field) values are prerequisites for the interpretation of magnetic anomalies and preparation of models. Following are the processes generally used for highlighting the contrast of magnetic anomalies.

4.14.1 REGIONAL–RESIDUAL SEPARATION

The most intriguing problem in dealing with the magnetic data is to separate out the regional and residual components. This can be achieved by adopting the following steps.

4.14.1.1 Wavelength Filtering

1. *High-pass filter.* The high-pass filter allows or passes all the high-frequency anomalies above certain cut-off values, representing shallow, narrow, and limited depth extent features. In other words, the process is to bring out the residual anomalies and remove the response of regional and deeper features.
2. *Low-pass filter.* This filter is applied when the user wants to remove the effect of shallow features and retain the response due to regional or deeper features.
3. *Band-pass filter.* As the name suggests, the filter passes a particular range of wavelengths. When the user wants to understand and evaluate the response of geological features situated at intermediate depths, the filter is applied.

4.14.1.2 Upward Continuation

The upward continuation technique is another tool to remove the responses arising from shallow sources.

- The magnetic anomaly grids can be continued upward by a distance, preferably the multiple of grid spacing, thereby rejecting the low-amplitude noise.
- Two or three such upward grids of increasing distance can be generated, and the one representing only the response of deeper features can be chosen to study the response of deeper features.
- The upward continued grid is removed from the original grid to obtain the residual grid. This is one way of regional–residual separation.

The upward continuation of the magnetic anomaly map at 2,000 m is presented in Figure 4.6. The nature of contours associated with high and low closures is trending NE–SW direction at this level of observation (Naskar and Lashkari, 2019). A prominent low zone identified in the map falls over the quartzite exposure around Amer. A high-magnetic-anomaly zone A–A' aligned in NE–SW direction encompassing isolated lenses of highs appears in the continued map. There is a definite break in amplitude south of Dausa which may be indicative of contact trending NNW–SSE direction and can be a favourable location for mineralization. The high magnetic closure at the central part, NW of Tunga, can be considered significant from a mineral point of view. This may be inferred as the causative sources at deeper level and is one body.

4.14.1.3 Trend Surface Analysis

Trend surface analysis is an analytical technique designed to separate observed data into regional and residual components. In this process, the user generates polynomials of different order representing surfaces of the anomaly under study and selects the one that best fits the regional surface for the given gravity or magnetic data. A particular polynomial that best fits will be taken for further qualitative and quantitative analysis of regional features. The value of the polynomial at each point is removed from the observed magnetic/gravity value to get the residual that best represents the area.

4.14.1.4 Downward Continuation

The downward continuation method helps enhance the responses due to shallow features which otherwise, in some cases, are masked under the influence of regional responses or gradients.

- The downward continued grid represents 'local anomaly' or residual anomaly, representing relatively shallow features, that is, the local features of interest.

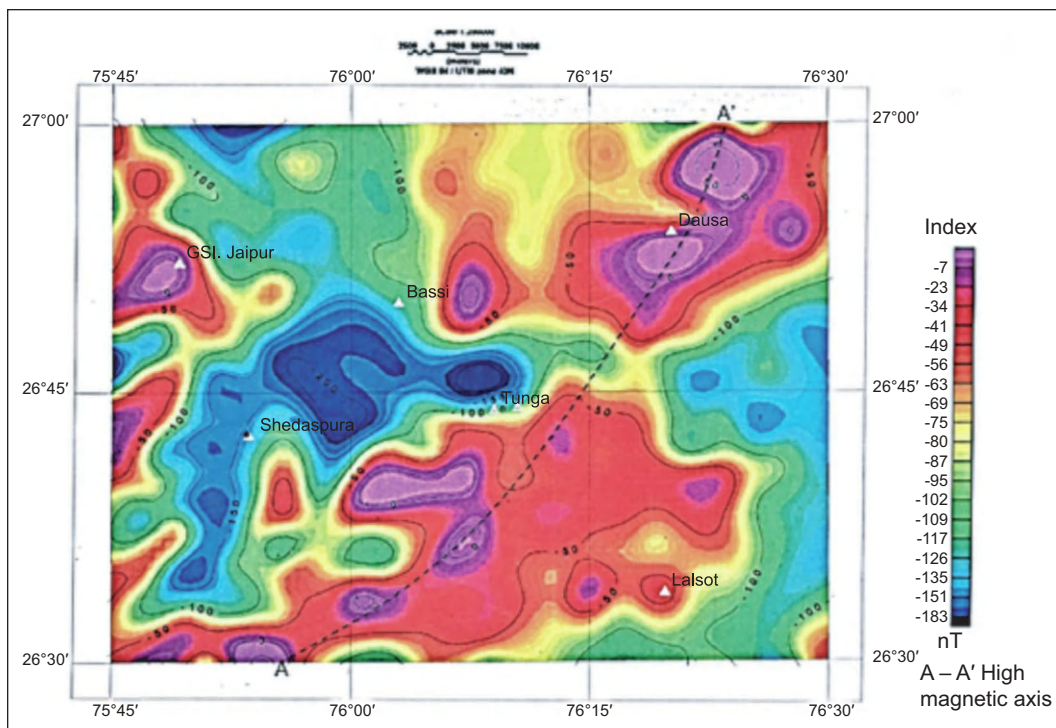


FIGURE 4.6 Upward continuation of magnetic anomaly map (2,000 m) of Toposheets 45 N/13, 14 and 54 B/1, 2, 5, 6 in Tonk and Dausa Districts, Rajasthan.

4.14.2 DERIVATIVES

Generally, the derivative grids pick up the subtle variations in magnetization contrasts which are otherwise masked or faintly reflected in the original grids. The derivatives can be computed along horizontal or vertical direction.

4.14.2.1 Horizontal Derivative Maps

Horizontal derivative is defined as the rate of change of the magnetic anomaly per unit distance in X or Y direction.

- These maps highlight the magnetization contrast and help bring out the lithological contacts or faults.

4.14.2.2 Vertical Derivative Maps

The *vertical derivative* is defined as the variation in the magnetization in vertical direction and helps in bringing out shallow features.

4.14.3 ANALYTIC SIGNAL

Analytic signal is a function related to all the three gradients of the magnetic field in X, Y, and Z directions and is given by the square root of the sum of the squares of the derivatives. This helps identify the edges of an anomalous body.

$$AS = \sqrt{(dx^2 + dy^2 + dz^2)}$$

where dx and dy are the horizontal derivatives and dz is the vertical derivative. The analytic signal for a profile data or spatial data is very effective in bringing out the boundaries of the causative sources.

The high-intensity magnetic anomaly closures between Regalagadda and Tummurukota, reflected as a zone of high analytic signal, may be attributed to the Peddavuru schist belt toward the western part (Figure 4.7; Naskar et al., 2020a). The analytic signal map of the magnetic data has brought out predominantly high-magnetic-anomaly peaks in the northern and eastern parts, which are also dominated by small bipolar anomalies in the original magnetic map (Naskar et al., 2020a).

4.14.4 REDUCED-TO-POLE (RTP)

Reduction to the pole (RTP) recalculates an observed total field magnetic anomaly data as if the inducing magnetic field had a 90° inclination. This transforms dipolar magnetic anomaly into monopolar anomalies centred on their causative bodies, thereby simplifying the interpretation. The symmetric 'highs' are directly centred on the body, while the maximum gradient of the antisymmetric dipolar anomalies coincides exactly with the body edge.

- Pole reduction is difficult at low magnetic latitudes (*around less than $21^\circ N$ latitude*), since north–south bodies have no detectable induced magnetic anomaly at zero geomagnetic inclination.
- Pole reduction is not a useful technique, where there are appreciable remanence effects.

A *reduction-to-pole* (RTP) correction shifts anomalies so that they are directly above the source bodies, like if the measurements took place at a pole, as seen in Figure 4.8. RTP corrections can be used for the correct location of source bodies, but the amplitude of the anomaly is unreliable (Ravat, 2007). On the left shows the magnetic anomaly imaged in the mid-latitudes; the anomaly appears to be off-centre from the source from the latitude. The right shows the anomaly directly above the source, after the reduced-to-pole correction is applied.

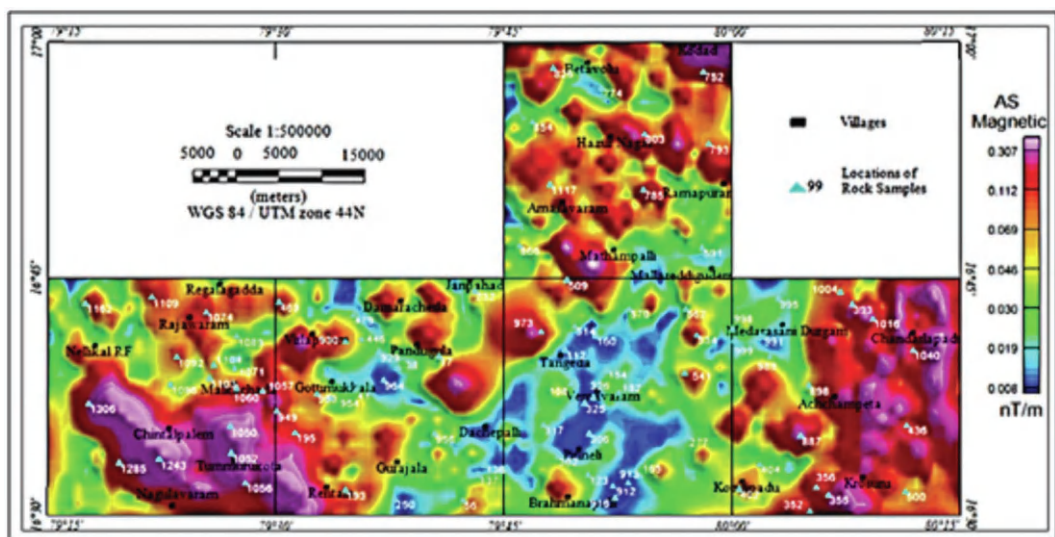


FIGURE 4.7 Analytic signal map of magnetic (TF) anomaly, parts of Khammam and Nalgonda Districts, Telangana.

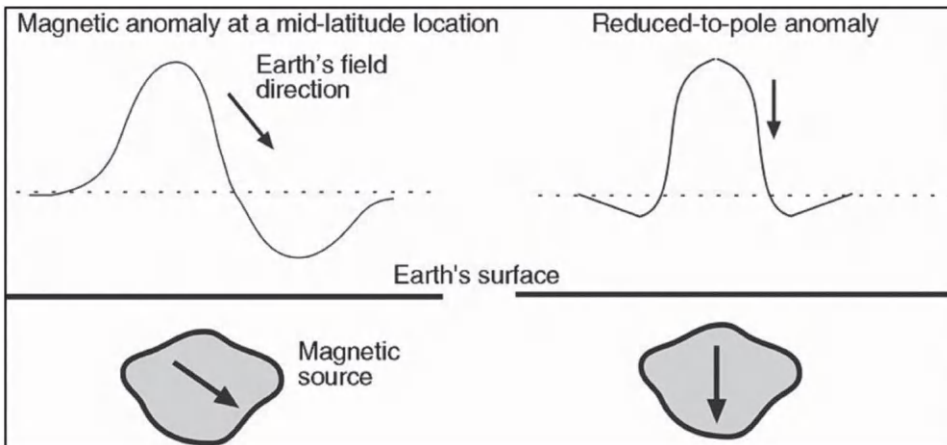


FIGURE 4.8 Image shows the effects of using reduction-to-pole correction.

4.14.5 REDUCED-TO-THE-EQUATOR (RTE)

The magnetic data obtained at latitudes less than 21°N latitude is not needed for producing RTE maps. However, favouring certain conditions, the RTE can be applied to the data collected at higher latitudes for qualitative analysis.

4.15 DATA INTERPRETATION

The interpretation of magnetic data is carried out in two stages, viz., the qualitative interpretation and the quantitative interpretation. The interpretation of magnetic data is complicated for the following reasons and hence must be carried out with caution.

- Magnetic anomalies are strongly influenced by the strike of the target and the latitude of its occurrence since the inducing field itself varies from equator to pole.
- Remanence.
- Assuming that the magnetization is caused purely by induction, for the north–south striking bodies at the equator, the Earth field magnetizes the body along its strike with no effective magnetization, and hence no magnetic anomalies.
- The magnetic method fails under these circumstances, in spite of the high-susceptibility contrast of the target.

4.15.1 QUALITATIVE INTERPRETATION

The qualitative interpretation of magnetic data following the techniques is required.

- Delineation of contour density and contour paucity zones.
- Linear gradient zones.
- Description of general trends, orthogonal trends, etc.
- Identification and marking of bipolar, high-dominant, and low-dominant anomaly zones to understand the nature of the causative source.
- Overall variation of magnetic anomaly in the study area (minimum and maximum values) should be recorded.
- Subtle magnetic variations (no signs) and expected signatures over typical ore environments, like alteration zones, etc., are to be identified.

- Magnetic discontinuities, shifts, and breaks in the anomaly pattern are to be recorded.
- Identification of fault zones characterized by breaks in the anomaly pattern and shifts in contouring are to be done.

4.15.2 QUANTITATIVE INTERPRETATION

The quantitative interpretation process begins by putting lines on maps, while the qualitative process is the start of the quantitative phase. Refinement of these locations begins with the determination of z , that is, depth values. For example, depth estimates to the tops of anomalous magnetic bodies are generated by a number of means, including slope measurement methods and analytic methods, such as Euler and Werner. Magnetic modelling including forward and inversion approaches contributes significantly to location in x , y , and z . Accurate results of all these rely upon sensible qualitative recognition of body types.

The 3-D Euler deconvolution of Reid et al. (1990) is a technique applied to the gridded potential field data to determine the position, depth, and nature of sources on the basis of the gradient of the potential field.

Depth solutions derived from magnetic anomaly have been plotted in the magnetic analytic signal map (Figure 4.9) for an example. Major responses are from source bodies occurring at depths of less than 2,000 m. In the southern part, maximum solutions of scattered distribution have been obtained for a depth of less than 1,000 m and lie over peaks of analytic signals, implying the depth of a magnetic interface in this area. This lends support to the power spectrum of magnetic anomaly, which shows a magnetic interface at 1.2 km depth. In other cases, linear distributions of plots of solutions for 1,000–2,000 m in southern and central parts of the area represent nearly ESE–WNW and ENE–WSW trending structural features within a 2 km depth (Naskar et al., 2020a).

4.16 PRESENTATION OF RESULTS (DELIVERABLES)

After all corrections have been made, deliverables typically include maps of the defined survey location, colour contour maps of the data acquired, residual–anomaly separation maps if necessary, and an interpretation map showing the cultural features and designated areas that are anomalous.

4.17 MAGNETIC ANOMALIES OVER SIMPLE SHAPES

Magnetic anomalies, on the other hand, are a function of two independent parameters:

- The subsurface distribution of susceptibility
- The orientation of the Earth's main magnetic field

The magnetic anomaly over a 2-D body such as a tunnel will look different depending on the orientation of the tunnel, say, east–west or north–south, even if the magnetic profile is always taken perpendicular to the trend of the tunnel.

Magnetic anomalies for an object of the same size, composition, and depth have different signatures at different magnetic latitudes because the magnetic inclination—the angle at which the magnetic force field is oriented with the Earth's surface—changes with latitude.

4.17.1 COMPARISON BETWEEN GRAVITY AND MAGNETIC ANOMALIES

Unlike the gravitational anomalies caused by subsurface density variations, the magnetic anomalies caused by subsurface variations in magnetic susceptibility are difficult to construct intuitively. This is because there are more factors involved in controlling the shape of a magnetic anomaly than there are in controlling the shape of a gravity anomaly.

distribution is moved to a different location on the Earth, say, from the equator to the North Pole. The gravity anomaly is a function of density distribution only.

Magnetic anomalies, on the other hand, are a function of two independent parameters: the sub-surface distribution of susceptibility and the orientation of the Earth's main magnetic field. Change one of these parameters and you change the resulting magnetic anomaly. What this means in practice is that magnetic anomalies over the same susceptibility distribution will be different if the distribution is in a different location, say, one located beneath the equator versus one located beneath the North Pole. Additionally, the magnetic anomaly over a two-dimensional body such as a tunnel will look different depending on the orientation of the tunnel, say, east–west or north–south, even if the magnetic profile is always taken perpendicular to the trend of the tunnel.

With these complexities in mind, we will not spend a great deal of time analysing the shapes of magnetic anomalies over simple structures; there are many computer programs available that do this quite well. Rather, we will look at several simple examples and qualitatively construct the magnetic anomalies over them so that you can get a better feeling for the complexities involved and for how it might be done in the computer.

4.17.2 MAGNETIC ANOMALY OVER MAGNETIZED SPHERE AT THE NORTHERN HEMISPHERE

Figures 4.10, 4.11, 4.12, 4.13, and 4.14 show profiles of magnetic total intensity anomalies for the same object at different latitudes in the Northern Hemisphere. In the Southern Hemisphere, the profiles would be the opposite (north and south would reverse south of the equator, and the inclination angles i would be negative).

4.17.2.1 Magnetic Anomaly over a Magnetized Sphere at the North Pole ($i = 90^\circ$)

Let us now qualitatively construct what the magnetic anomaly of a metallic sphere located beneath the North Pole would look like. The geometry of the sphere, the Earth's main magnetic field, the field lines associated with the anomalous field, the direction and magnitude of the anomalous field, and a plot of the intensity of the anomalous field that would be recorded are shown in Figure 4.10.

At the North (magnetic) Pole, the Earth's main magnetic field, F_e points straight down. Because the buried sphere is composed of a material with a non-zero susceptibility, the Earth's main magnetic field causes the sphere to produce an induced magnetic field. Field lines associated with this induced field are shown by black lines, and the magnitude and direction of the induced, anomalous field, F_{ai} , at the surface of the Earth are shown by the blue arrows.

The total field, whose strength will be recorded on a proton precession magnetometer, will be the sum of the main field, F_e , and the induced, anomalous field, F_{ai} . Notice that to either side of the sphere, the anomalous field points in the opposite direction as the main field. Thus, when the main field is removed from our observations, we will observe negative values for the anomalous field. Near the sphere, the anomalous field points in the same direction as the main field. Therefore, when the main field is removed, we will observe positive values for the anomalous field.

4.17.2.2 Magnetic Anomaly over a Magnetized Sphere at the Equator ($i = 0^\circ$)

At the equator (magnetic), the direction of the Earth's main magnetic field is now horizontal. It still induces an anomalous magnetic field in the metallic sphere, but the orientation of field lines describing the magnetic field are now rotated 90° . The field lines are indicated by the black lines, and the strength and direction of the anomalous field at the surface of the Earth are shown by the blue arrows (Figure 4.11).

Above the sphere, the anomalous magnetic field, F_{ai} , now points in the opposite direction as the Earth's main magnetic field, F_e . Therefore, the total field measured will be less than the Earth's main field, and so upon removal of the main field, the resulting anomalous field will be negative (Earth Science Department University of Melbourne, Australia). On either side of the sphere, the anomalous field points in the general direction of the main field and thus reinforce it, resulting in total field measurements that are larger than the Earth's main field. Upon removal of the main field contribution, these areas will show positive magnetic anomalies.

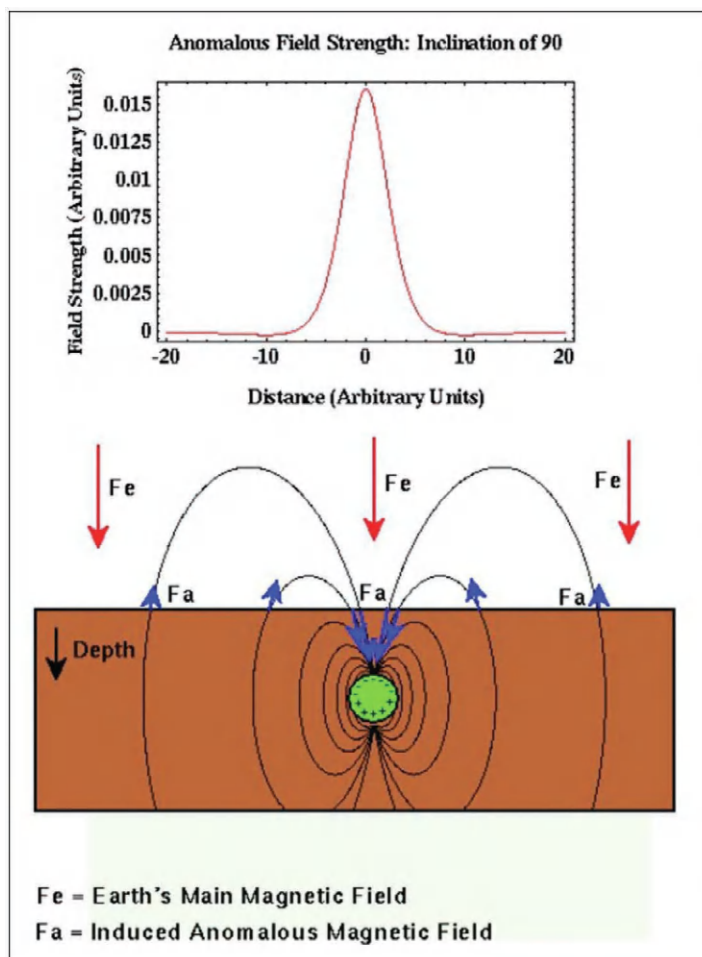


FIGURE 4.10 Magnetized sphere at the Earth's North Pole.

Source: Earth Science Department University of Melbourne, Australia.

4.17.2.3 Magnetic Anomaly: Magnetized Sphere in the Northern Hemisphere ($i = 45^\circ$)

Finally, let us examine the shape of the anomalous magnetic field for a metallic sphere buried somewhere in the Northern Hemisphere, say, near Denver. As in the previous examples, the Earth's main magnetic field induces an anomalous field surrounding the sphere. The anomalous field is now oriented at some angle, in this case 45° , from the horizontal. By looking at the direction of the anomalous field, F_a , in comparison with the Earth's main field, F_e , you can see that there will be a small negative anomaly far to the south (to the right in the diagram) of the sphere, a large positive anomaly just south of the sphere, and a small broad negative anomaly north of the sphere. Notice that the magnetic anomaly produced is no longer symmetric about the sphere (Figure 4.12). Unless you are working in one of those special places, like the magnetic poles or equator, this will always be true (Earth Science Department University of Melbourne, Australia).

We can consider the situation in the Southern Hemisphere, at the same magnetic latitude, by making some mental adjustments to the same diagram. If we take north to be to the right and reverse the direction of the main field arrows, we have created the Southern Hemisphere setting. The shape of the induced, anomalous field will be the same, but the direction will reverse as a consequence. The

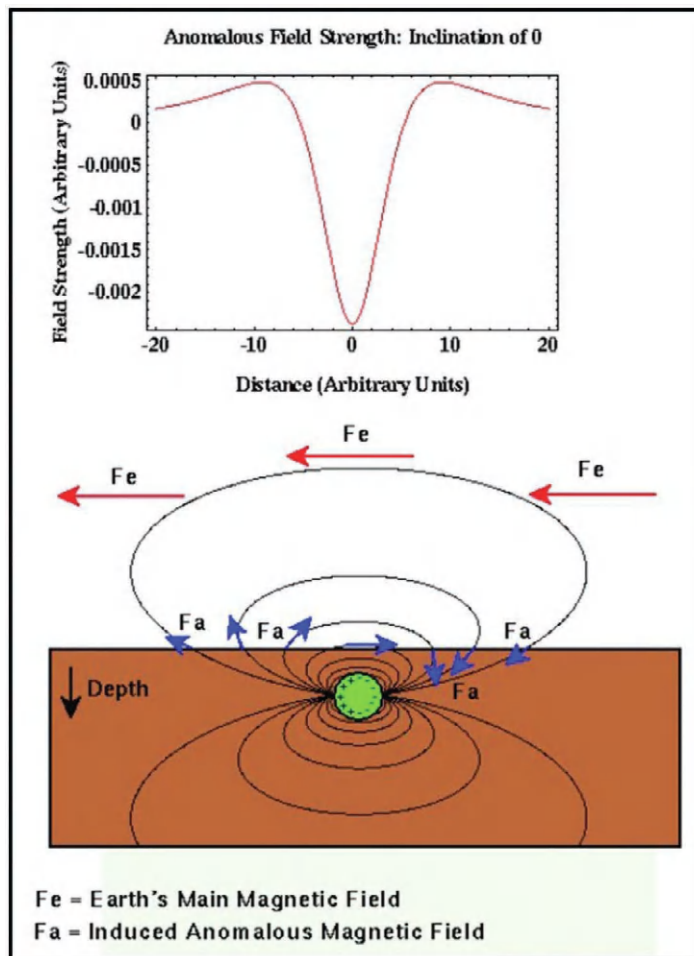


FIGURE 4.11 Magnetized sphere at the Earth Equator.

outcome is that the profile will have exactly the same shape! A generalization is that, where there is an isolated anomaly due to magnetization by induction, the maximum will always be on the equator side of the anomaly, and the minimum will be on the pole side.

4.17.2.4 Magnetic Anomaly: Magnetized Sphere When $i = 67\frac{1}{2}^\circ$

The anomalous field is oriented at an angle relative to the horizontal plane. In the example, the inclination angle (i) is $67\frac{1}{2}^\circ$. This means that the magnetic field lines around the sphere deviate from the horizontal direction by $67\frac{1}{2}^\circ$. Unlike gravity anomalies, which tend to be symmetric around the source, magnetic anomalies exhibit more complex shapes. The magnetic anomaly produced by the sphere is no longer symmetric about the sphere's centre. Specifically, we observe the following features: a small negative anomaly far to the south of the sphere; a large positive anomaly just south of the sphere; a small, broad, negative anomaly north of the sphere. Visual interpretation of magnetic anomalies can be challenging due to their non-symmetrical nature. However, computer modelling algorithms can accurately simulate and interpret magnetic anomalies. By incorporating the survey location and other relevant parameters, these algorithms generate appropriate magnetic models.

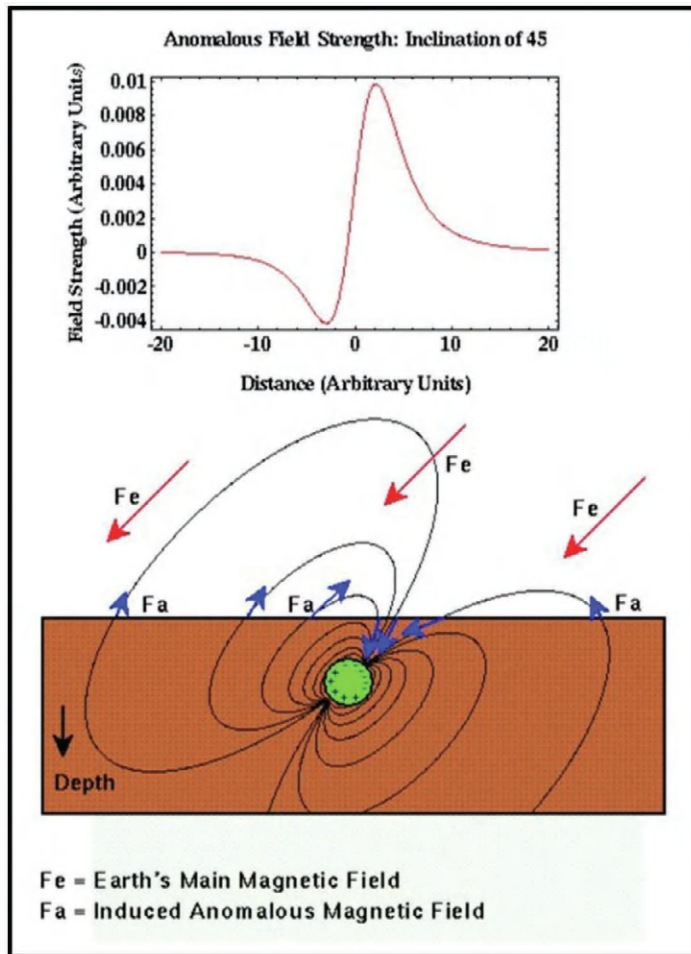


FIGURE 4.12 Magnetized sphere in the Northern Hemisphere.

4.17.2.5 Magnetic Anomaly: Magnetized Sphere When $i = 22\frac{1}{2}^\circ$

Imagine a magnetized sphere (such as a magnetic rock or mineral) located within the Earth's crust. When a magnetized sphere (modeled as a dipole) is located at an inclination (also known as the dip angle) $i = 22\frac{1}{2}^\circ$, it means that the magnetic field lines within the sphere are tilted downward from the horizontal plane by that angle. In this case, the magnetic field lines emanating from the sphere are inclined at $22\frac{1}{2}^\circ$. In summary, the inclination of a magnetized sphere at $22\frac{1}{2}^\circ$ can indeed lead to magnetic anomalies, and whether they are positive or negative depends on the alignment of the sphere's magnetization relative to the Earth's magnetic field. Geophysical studies and modelling techniques help us understand these anomalies and infer geological properties based on magnetic data.

From this simple set of examples, you now see that it is indeed more difficult to visually interpret magnetic anomalies than gravity anomalies. These visual problems, however, present no problem for the computer modelling algorithms used to model magnetic anomalies. You simply need to incorporate the location of your survey into the modelling algorithm to generate an appropriate magnetic model.

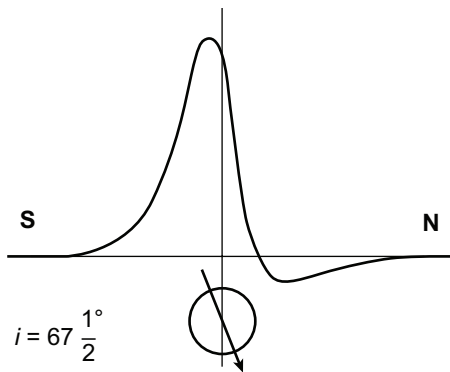


FIGURE 4.13 Magnetized sphere when $i = 67 \frac{1}{2}^\circ$.

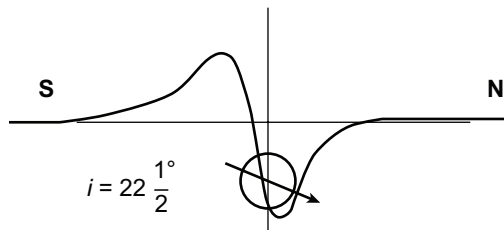


FIGURE 4.14 Magnetized sphere when $i = 22 \frac{1}{2}^\circ$.

Source: Wikipedia.

4.18 APPLICATIONS OF MAGNETIC SURVEYING

Magnetic surveying equates to the search for magnetite. Magnetite tracers often disclose the location of other minerals. The objective is to determine the depth and structure of mineral-bearing bodies (Naskar and Lashkari, 2019; Naskar et al., 2020b). Other minerals that may be prospected with magnetic surveying include nickel, copper, and gold, because these are often associated with magnetite (Naskar et al., 2018, Naskar and Kulkarni, 2019). Kimberlite is an ultrabasic rock that contains magnetite and ilmenite and may thus be detected by magnetic surveying (Das et al., 2005). Magnetic surveying may be useful in basalt-covered areas, where seismic reflection will not work, for example, the Columbia River Basalt province, Washington State, USA, and the Central Indian Shield (CIS). There, the approach used is to combine magnetotelluric measurements, which are powerful to give the vertical structure, with gravity and magnetics, which are powerful to give lateral variations (Naskar et al., 2008).

Up to quite recently, magnetic surveying was used to determine the depth to the basement only (i.e. it was not much use at all). However, this was, in general, accurate to ~5%. Magnetic results can also show topographic trends in the basement that may be relevant to the structure in the sediments above. Some oilfields have been discovered almost entirely because of the magnetic field, usually if the production is from porous serpentinite, which has a high susceptibility compared with surrounding sediments. Sometimes oil exists in sediments draped over a basement high (Naskar et al., 2016).

The search for sediment is a new field and is leading to a resurgence in magnetic work. It has been made possible because of the advent of high-resolution/high-sensitivity surveys. The

Gondwana sediments have been deposited in several known palaeochannels (faulted basins) over the Precambrian rocks in India. These sediments are known to be coal-bearing and mined from these basins in the states of West Bengal, Bihar, Madhya Pradesh, and Maharashtra. A large lava flow of basalt covered the Gondwana and Triassic sediments during the Cretaceous period, which is known as the Deccan trap. Subsequently, tertiary rocks and older alluvium covered this flow at many places. The area was then subjected to tectonic disturbances, which resulted in the formation of faulted basins.

4.18.1 CASE STUDY: CENTRAL INDIA

The Sohagpur, Central India, and Wardha Valley coalfield areas were covered by the magnetic survey in order to delineate the Gondwana sediments below the trap cover (Naskar and Rai, 2018; Naskar et al., 2003, Figure 4.15; Naskar and Saha, 2015). Magnetic survey along the 350 km long Ujjain–Buldana profile exhibits two fault zones in the Malwa Plateau and one fault zone each in the Chhegaon–Makhan and Borgaon–Burhanpur areas. These are corroborated by the findings of the gravity survey. South of Tapti River the broad magnetic low indicates a thin cover of Deccan trap overlying non-magnetic Gondwana sediments. The resultant basement configuration by 2-D modelling clearly shows that Deccan traps are comparatively thin in the Narmada River region. The interface between the Deccan trap and the underlying formation is interpreted at a depth of 600 m around Sanwad, which is corroborated with resistivity data (Figure 4.15).

4.18.2 CASE STUDY: ARCHAEOLOGICAL RESEARCH FOR LANDSCAPE STUDIES, KREUTTAL, AUSTRIA

The focus lies on the application of geophysical methods, especially magnetometry in archaeological research, for landscape studies in 20 km to the north of Vienna, Austria, containing a multi-period landscape. The aim is to study the landscape using noninvasive methods, including remote sensing, geophysical prospection, virtual reality, and GIS (<http://lbi-archpro.org/cs/kreuttal/>).

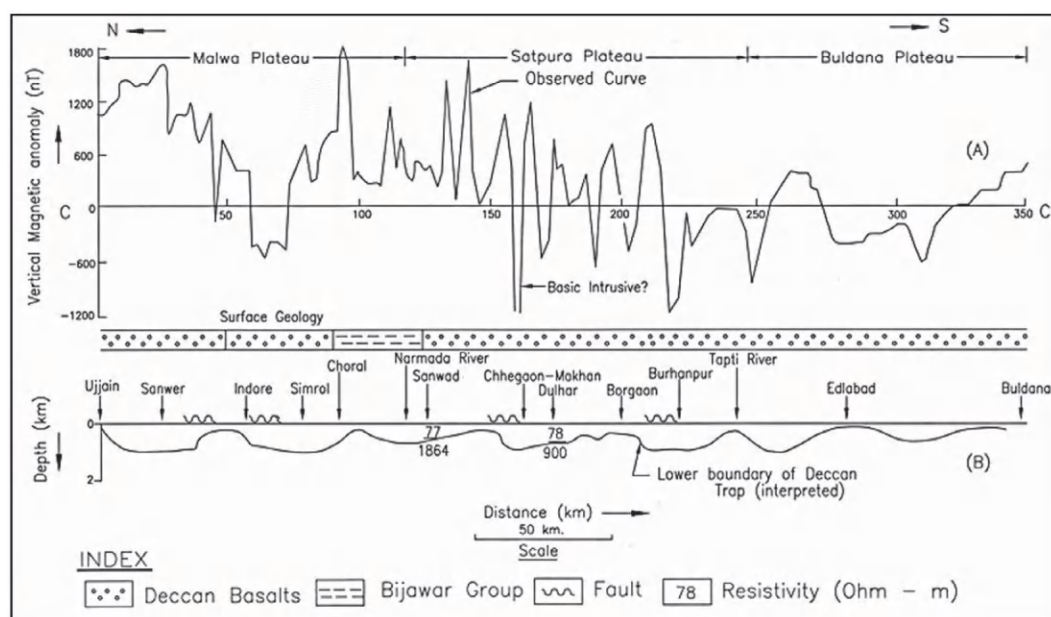


FIGURE 4.15 Lower boundary of the Deccan trap is delineated by 2-D magnetic modelling.

A motorized magnetometry survey of ~4 km² was collected by the LBI Arch Pro using eight flux-gate sensors with a measuring raster of 25 cm × 12.5 cm (Kainz, 2019). Vertical photographs were used to complement the magnetometry data, while airborne laser scanner (ALS) data provided the topographic background. The datasets were collated within a GIS system and used for the initial mapping of features and structures.

Small area magnetometry and magnetic susceptibility (MS) surveys were undertaken prior to and during excavations. Firstly, this helped improve the magnetic definition of features before and after topsoil removal. Secondly, it enabled the identification of new features, especially features not visible to the eye during the excavation and topsoil magnetometry surveys. This enabled an assessment of feature visibility, especially the impact of the topsoil and other deposits, highlighting the biases and deficiencies in the topsoil magnetometry data (Kainz, 2016; Kainz and Cotter, 2018). An example is a palisade with postholes that appeared only partially in the magnetometry data; the full extent was only visible magnetically after the removal of the topsoil, and the postholes were only identifiable in the MS data (Kainz, 2019).

Chi-square tests were used to identify relationships within the data by creating analytic units to identify nonrandom associations between attributes of artefacts and emic traits that are intended and consistently produced (VanPool and Leonard, 2011). These attributes are related to the physical properties of features and structures and their occurrence throughout the landscape. Structures (e.g. houses) and features (e.g. pits, ditches, postholes, etc.) were analyzed for their quantity, magnetic values, orientation, size in relationship to their location (Kainz, 2019).

The mapping of the magnetometry data within a GIS system enabled the application of the research framework. The geophysical surveys during the excavations provided a qualitative assessment of the magnetometry data addressing the invisibility of certain features. Features within and between sites were investigated, allowing for a comparison between them. The analysis raised new questions for further research, including why sites show different characteristics in the magnetometry data and in variations in the magnetic properties, quantity, size, orientation of features, and structures. This allowed insights to be gained for the preservation and distribution of archaeological remains throughout the landscape.

4.18.3 CASE STUDY, SWITZERLAND: LIMITATIONS OF GEOTECHNICAL-GEOPHYSICS NON-UNIQUENESS

The most significant limitation of geotechnical geophysics is its non-uniqueness. In the absence of any external constraints (ground truth or basic conceptual model), a single geophysical dataset can be transformed into an infinite number of ‘theoretically correct’ output models (Table 2.1, Chapter 2). Figure 4.16 (Anderson et al., 2008), for example, shows magnetic data acquired in Switzerland and a corresponding subsurface interpretation that is theoretically accurate (consistent with field data), but not likely geologically consistent, since geologic features are not commonly shaped like question marks, as is the case in Figure 4.16. Of the numerous theoretical interpretations that could be generated for this magnetic dataset, the most reasonable model is the one that is most consistent with all other geophysical datasets and available ground truth.

4.18.4 CASE STUDY: NICKEL SULPHIDE EXPLORATION IN THE KAMBALDA DISTRICT, WESTERN AUSTRALIA

As noted by Pridmore et al. (1984), potentially all the common geophysical techniques have applications in direct exploration for nickel sulphides: gravity for the detection of dense massive and matrix ore, magnetics for the detection of magnetite and pyrrhotite in the ores, electromagnetics for the detection of the massive and matrix ore, and induced polarization for the detection of disseminated and, to some extent, matrix and massive ore. However, Trench and Williams (2019) draw attention

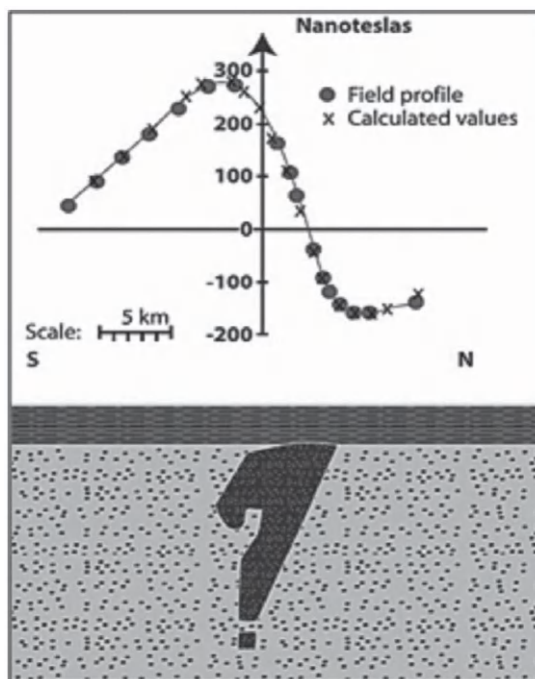


FIGURE 4.16 Magnetic data acquired in Switzerland, and a corresponding subsurface interpretation that is ‘theoretically accurate’ but geologically absurd.

to the significant limitations imposed by the small size of the deposits, the extreme depth of weathering on the Yilgarn Craton, and the abundance of geological noise or anomalous responses caused by non-economic sources. Nevertheless, even ‘blind’ nickel ores can still give rise to recognizable geophysical anomalies (Trench and Williams, 2019, Figure 3, p. 171; ASEG Extended Abstracts, 1994, Taylor & Francis).

The difficulties posed for geophysical exploration in the Yilgarn Craton have been extensively documented elsewhere (Doyle et al., 1981; Doyle and Lindeman, 1985). The major obstacles to successful exploration are summarized as:

1. There is a thick, conductive regolith which locally exhibits a strong magnetization. In fact, there are significant variations in thickness–conductivity and magnetization both laterally and vertically within the regolith.
2. A comparable conductivity–thickness product between nickel sulphides and barren sedimentary units. In the Kambalda region, problems are compounded by the substantial thicknesses (locally up to 90 m) of conductive salt lake sediments in many areas.

From a positive viewpoint, if high-resolution ground magnetic surveys are employed to characterize fully the magnetic signature of surficial laterites, frequency and non-linear filtering effectively remove their influence while maintaining considerably more detail than aeromagnetic data (Stanley et al., 1992). Conversely, the regolith itself may contain very useful information pertaining to the distribution of subsurface rock types and mineralization. For example, the ultramafic lavas which stratigraphically overlie the nickel mineralization generally weather more deeply than do the ‘footwall’ mafic lavas and, thus, produce a more conductive, thicker weathering profile. The latter may be mappable using a variety of techniques, such as EM, magnetics, gravity, seismic refraction, and shallow seismic reflection.

4.18.5 MAGNETIC ANOMALY ABOVE AN OLIVINE DIABASE DYKE, PISHABO LAKE, CANADA: CASE STUDY

From the point of view of geophysicists, geophysical methods play an important role in mineral investigation. The magnetic method is one of the oldest branches of geophysics and used in many exploration issues, such as mineral and ores as massive sulphide, iron, gold, and porphyry copper deposits. Magnetic data interpretation has shown its efficiency in the identification of deep and shallow structures known to employ a structural control on mineralization occurrences (Essa et al., 2018; Le Maire and Munsch, 2018). Uniform geological models (geologic contacts, thin sheets, cylinders, and spheres) are frequently employed in magnetic inversion to estimate the body factors (the amplitude factor, the depth, the index angle, the location of the origin, and the shape) and have a vital role in many exploration issues. These models cannot be an exact geologically representation but are generally a good tool in magnetic interpretation to calculate, in particular, the body parameters. Figure 4.17 demonstrates the magnetic anomaly profile of length 2,200 m above an olivine diabase dyke from the Pishabo Lake, Canada, and this site is made out of plagioclase, purplish-brown augite, pale-green olivine, apatite, some biotite, and large patches of magnetite (McGrath and Hood, 1970).

This magnetic anomaly has been interpreted by using various inversion algorithms, such as the moving average method, the parametric inversion method, and the PSO method. The elucidation procedure and their produced results are mentioned in Table 4.4. The predicted parameters (M , which represents the amplitude factor; z , the depth; θ , the magnetization angle; d , the origin location; and q , the shape) of the body by using these inversion methods have a good agreement together.

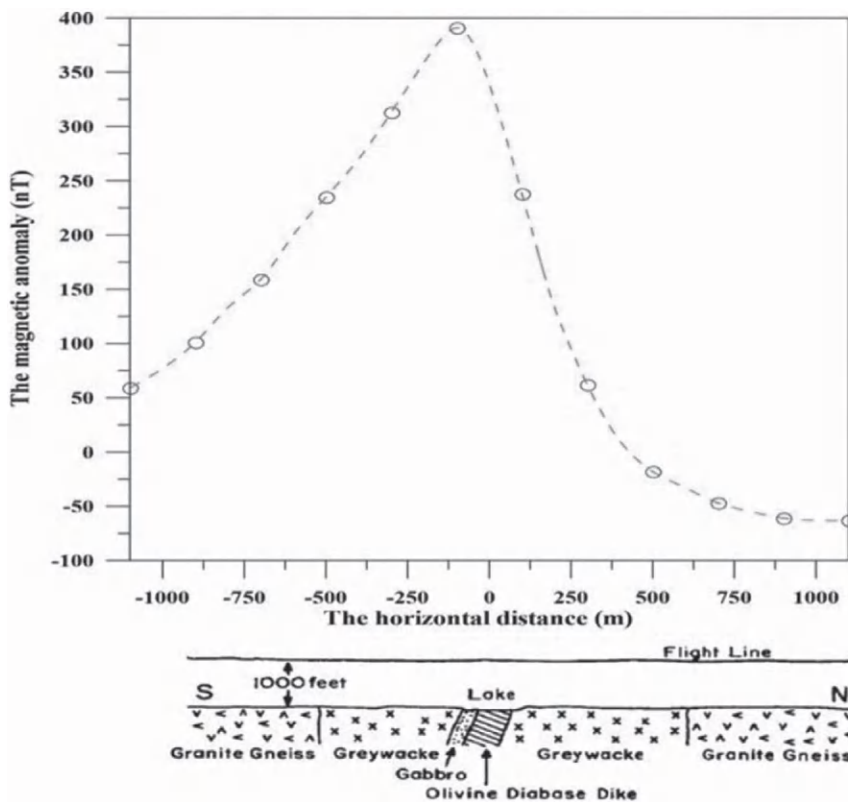


FIGURE 4.17 The magnetic anomaly over an olivine diabase dyke from the Pishabo Lake, Canada.

Source: Adapted from Essa and Munsch (2019).

TABLE 4.4

Results Obtained for Interpreting Magnetic Anomaly over an Olivine Diabase Dyke from the Pishabo Lake, Canada

Parameters	McGrath and Hood (1970) Method	Abdelrahman et al. (2009) Method	Abdelrahman et al. (2012) Method	Biswas (2016) Method	Essa (2011) Method
M (nT)	—	—	1,429.0	1,411.8	1,380.8
z (m)	304.0	318.9	320.0	324.0	305.9
θ (°)	—	—	37.5	−37.9	38.5
d (m)	—	2.86	—	1.7	2.22
q (dimensionless)	—	1.0	1.0	1.0	0.98

Source: Essa and Munsch (2019).

4.18.6 MAGNETIC METHOD FOR ARCHAEOLOGICAL SITE CHARACTERIZATION OF THE JORDAN AREA, WESTERN ASIA: CASE STUDY

The magnetic method is the most frequently used geophysical tool for archaeological prospection, because of its speed and resolution in mapping large areas. It was first used in 1957 in England (Belshe, 1957, Aitken et al., 1958) and has since become the backbone of archaeological prospection. It has now been used even more frequently than electrical prospection methods. This method is used to map buried stone foundations and to outline the locations of forges and kilns, hearths, and campfire sites (Gibson, 1986). The utility of magnetic susceptibility measurements of topsoil in the mapping of anthrosoils (soils modified by human activity) was described by Colani and Aitken (1966). Recent advances in technology led to acquiring a large database and more precise evaluation of field data (Weymouth, 1986; Scollar, 1962; Al-Zoubi et al., 1998; Batayneh et al., 2001, 2006).

Archaeomagnetism (Tite and Mullins, 1971; Tarling et al., 1986) is a developing science that seeks to obtain site occupation dates (dating method) from the orientation of the remnant magnetic field found in stones lining ancient kilns. *Archaeomagnetism* is a nonprospecting method. Recent summaries of this specialty are available in Wolfman (1984) and Tarling et al. (1986). Isotopic analysis is another nonprospecting method. It is used both for dating purposes (for example, radiocarbon dating) and as well as for provenance studies. The figure shows a 3-D map of high-pass filtered total magnetic field over the Nabataean Hawar archaeological site in southern Jordan. The magnetic data shown in Batayneh (2011) indicates a fluctuation over the visible rectangle (cistern), while there is only negligible scatter in the part of the map outside the cistern walls. The cistern was 16 m east to west by 26 m north to south. Because of these significant magnetic changes, the general location and dimensions of the cistern were determined.

4.18.7 GEOSPATIAL ARCHAEOLOGY, LOTHAL REVISITATION PROJECT, GUJARAT, INDIA: CASE STUDY

This project aimed to detect different natural and artificial subsoil features, complemented by a series of core drillings to determine the shifting of palaeochannels and shorelines, to reconstruct the palaeogeography around Lothal during the Late Mid-Holocene and the hydraulic structures that interfaced the site with the surrounding environment (Frenez, 2014). Lothal is a well-known Harappan civilization site with evidence of occupation during the second half of the 3rd millennia BCE (Prabhakar and Korisettar, 2017).

This site has revealed the presence of a twin fold habitation surrounded by fortification, craft-activity areas, a warehouse, maritime trade and external contacts, cemetery, and a dockyard. The

techniques used in the investigations include geomorphological, remote sensing analysis of the area around Lothal, a 3-D digital elevation model using relative kinetic GPS, magnetic survey of the non-excavated archaeological area using caesium magnetometer. The magnetometer survey showed three prominent anomalies: (1) baked brick embankment canal running east–west and perpendicular to the dockyard, probably connecting the nearby palaeoriver to the dockyard (anomaly A); (2) an architectural complex consisting of rooms along a narrow street and separated by lanes to the north-eastern area (anomaly B); and (3) a possible monumental gateway or a large drainage outlet to the south-western corner of the acropolis (anomaly C). Frenez (2014) also carried out ‘test trenches’ in the areas of anomalies, which largely confirmed with the findings (Prabhakar and Korisettar, 2017). Thus, application of noninvasive techniques of unexcavated areas at Lothal had reduced not only the time spent in speculative excavations to find buried features but, rather, also using the geophysical techniques to first understand the subsoil features and then proceeding to limited excavations with maximum output.

4.19 GEOMAGNETIC DATA

Chapman and Bartels (1940) and Vestine (1960) represented the source of the regular magnetic variation in terms of a current sheet at an altitude of ~100 km above the Earth’s surface with two vertices, at around 350° latitude. Regular and systematic variations of the magnetic components at many observatories around the globe have subsequently brought out an important finding by way of the equatorial electrojet, a phenomenon representing the enhanced flow of east–west current over the magnetic equator. The geographical location of India in the global scenario is ideal for monitoring the developments in the equatorial electrojet activities and its associated effects in the global current system. The existing extensive network of magnetic observatories through Russia from the magnetic pole to Peninsular India centred along the 145° geomagnetic meridian provides unique datasets of magnetic variations from pole to dip equator. Data from magnetic observatories are utilized to study the time-varying phenomenon and its spatial variability.

4.19.1 GEOMAGNETIC OBSERVATORIES IN INDIA

The Indian Institute of Geomagnetism (IIG) (www.iigm.res.in) is currently operating 12 strategically located magnetic observatories covering the length and breadth of India. Each magnetic observatory is set up with a specific scientific interest of investigation, and various scientific results are being published by IIG. Geomagnetic signals generated or propagated through the ionospheric region get modified at the dip equator, that is, geomagnetic field variations undergo equatorial enhancement during daytime. The Tirunelveli (TIR) Magnetic Observatory (MO), located in this equatorial region, started operations in 1996. Pondicherry MO (PND; 1993) is located in the fringe region of the equatorial electrojet belt. The prime MO Alibag (ABG; 1904), successor to the Colaba Observatory (1826–1904), is located on the western coast of India, away from the influence of the equatorial electrojet. IIG India is holding the legacy of long geomagnetic data series. ABG (Tiwari et al., 2014) is a calibration centre for magnetic compasses of Indian Navy and Air Force and serves as a base station for marine magnetic surveys carried out off the Arabian Sea coast.

Magnetic storm days are classified from magnetograms of Indian stations. The details of currently operating magnetic observatories are provided in Table 4.5. In addition to all these magnetic observatories (Table 4.5) operated by IIG, two more magnetic observatories in India, namely, Hyderabad (HYD) and Sabarwala (SAB), are operated by CSIR National Geophysical Research Institute, Hyderabad, and Survey of India, Dehra Dun, respectively, and both started functioning in 1964. HYD MO initiated variometer measurements using the LaCourVariometer (HZD) instrument and later upgraded to modern magnetic observatory and became an intermagnet observatory since 2009. A new magnetic observatory is being set up by CSIR-NGRI at Choutupal (CPL) with digital measurements, and it will be a successor to HYD; locations of the observatories are plotted (Chapter 14; Figures 14.2–3).

TABLE 4.5**Magnetic Observatories in India**

Magnetic Observatory	IAGA Code (Operating Institute)	Instrument Type	Start	Geographic		Geomagnetic	
				Latitude	Longitude	Latitude	Longitude
Tirunelveli	TIR (IIG)	Izmiran	1996	8°42' N	77°48' E	0.15° N	150.71° E
		DFM	2000				
Port Blair	PBR (IIG)	DFM	2011	11°48'	92°43'	2.28°	165.56°
Pondicherry	PND (IIG)	Izmiran	1993	11°55'	79°55'	3.16°	153.05°
		DFM	2000				
Visakhapatnam	VSK (IIG)	Izmiran	1994	17°4'1	83°19'	8.64°	156.77°
		DFM	2001				
Alibag	ABG (IIG)	Izmiran	1975	18°37'	72°52'	10.40°	146.81°
		DFM	1993				
Nagpur	NGP (IIG)	Izmiran	1991	21°09'	79°05'	12.38°	152.99°
		DFM	2003				
Rajkot	RKT (IIG)	DFM	2007	22°18'	76°47'	13.70°	150.90°
Silchar	SIL (IIG)	Izmiran	1999	24°56'	92°49'	15.33°	166.23°
		DFM	2011				
Allahabad	ALH (IIG)	DFM	2005	25°27'	81°51'	16.44°	155.96°
Shillong	SHL (IIG)	Izmiran	1975	25°34'	91°53'	16.00°	165.39°
		DFM	2012				
Jaipur	JAI (IIG)	Izmiran	1975	26°55'	75°48'	18.36°	150.41°
		DFM	2002				
Gulmarg	GUL (IIG)	Izmiran	1977	34°05'	74°24'	25.59°	149.87°
		DFM	2005				
Hyderabad	HYD (NGRI)	Izmiran	1964	17°25'	78°33'	8.66°	151.88°
		DFM	2008				
Sabhawala	SAB (SOI)	Askania	1964	30°22'	77°48'	21.62°	152.37°

Note: IAGA code and operating institute names are also mentioned in column 3. *Source:* IIG, NGRI, SOI.

Data from all the magnetic observatories are processed with suitable software, and the digital data are compared between stations and checked for spikes, artificial disturbances, noise, jumps, time shifts, etc. The raw data are then despiked; artificial jumps, if any, are adjusted; and 3 minute running averages are carried out on noisy data to reduce the scatter. Also, any loss of data in one system is replaced by the data from the alternate magnetometer system by adding appropriate constants to the variation data such that the data for consecutive days have continuity.

4.19.1.1 Intermagnet Network

The International Association of Geomagnetism and Aeronomy (IAGA) initiated the Intermagnet network to exchange online data from important magnetic observatories under one umbrella during 1987. Intermagnet (www.intermagnet.org) is the global network of geomagnetic observatories that monitors the Earth's magnetic field and helps in adopting modern standard specifications for measuring and recording equipment in order to facilitate data exchanges in close to real time. Observatories send 1 minute data in near real time using standard Intermagnet formats via satellite, computer, and telephone networks to geomagnetic information nodes (GINs). It is an accreditation to magnetic observatory and maintenance of high standards of data quality. ABG since 1997, HYD from 2009, and recently, Jaipur magnetic observatory (2013) are members of the Intermagnet network in India.

4.19.1.2 Magnetic Anomaly Maps

Magnetic measurements made over specific regions by different organization are sources of magnetic anomaly maps in India. These data do not cover the entire country and had been acquired on the ground and by aircraft at different altitudes. An attempt is made to compile all the available data and prepare a map at constant altitude, that is, 5,000 ft. The digital data as well as old processed maps can be procured from the Geological Survey of India (GSI) (www.gsi.gov) on nominal charges. Obtaining magnetic records of geomagnetic observatories is straightforward, and data can be acquired on request. For some stations, which are part of the international network, data can simply be downloaded. Using hourly and 1 minute geomagnetic data from observatories of the equatorial station Tirunelveli and the low-latitude station Alibag, Equatorial Electrojet Index (EEJ) are prepared periodically.

NOTES

- 1 The *expression potential field* refers to a mathematical property of these types of force fields. Both gravitational and magnetic forces are known as conservative forces. This property relates to work being path-independent. That is, it takes the same amount of work to move a mass, in some external gravitational field, from one point to another, regardless of the path taken between the two points. Conservative forces can be represented mathematically by simple scalar expressions known as potentials. Hence the expression potential field.
- 2 One order of magnitude is a factor of ten. Thus, four orders of magnitude represent a variation of 10,000.
- 3 By this we are only referring to that portion of the gravity field produced by the internal density distribution and not that produced by the tidal or drift components of the observed field. That portion of the magnetic field relating to internal Earth structure can vary significantly with time.
- 4 For plotting purposes, the arrow lengths shown in the earlier figures decay proportional to one over the distance between the two poles rather than proportional to one over the square of the distance between the two poles. If the true distance relationship were used, the lengths of the arrows would decrease so rapidly with distance that it would be difficult to visualize the distance–force relationship being described.
- 5 If we were to plot the force of gravitational attraction between two point masses, the plot would look identical to this.
- 6 Unlike the force plots shown on the previous page, the arrows representing the force have not been rescaled. Thus, you can now see how rapidly the size of the force decreases with distance from the dipole. Small forces are represented only by an arrowhead that is constant in size. In addition, please note that the vertical axis in the preceding plot covers a distance almost three times as large as the horizontal axis.
- 7 We have not mentioned previously that the source of all magnetic fields is most probably circulating electric currents—the field, due to a current loop, is identical with the field due to a dipole pair. Orbiting electrons in an atom are therefore conceptually responsible for a magnetic field which can react with the ambient field or which can be acted on by the ambient field.
- 8 Although susceptibility is unitless, its values differ depending on the unit system used to quantify H and I . The values given here assume the use of the SI, International System of Units (Système International d’Unités), based on the meter, kilogram, and second. Another, older unit system, the cgs—centimeter, gram, and second system—is also used in exploration geophysics. To convert the SI units for susceptibility given earlier to cgs, divide by 4.
- 9 We will assume that there is no remnant magnetization throughout the remainder of this discussion.
- 10 Unlike gravity, magnetic anomalies are rather complex in shape, and making sweeping statements like this can be very dangerous.
- 11 *Magnetic saturation* refers to the induced magnetic field produced in the bars. In general, as the magnitude of the inducing field increases, the magnitude of the induced field increases in the same proportion relating the external to the induced magnetic fields. For large external field strengths, however, this simple relationship between the inducing and the induced field no longer holds. Saturation occurs when increases in the strength of the inducing field no longer produce larger induced fields.
- 12 *Precession* is the motion-like experience of a top as it spins and ‘wobbles’. Because of the Earth’s gravitational field, not only does a top whose spin axis is not parallel to the direction of gravity spin about its

axis of rotation but the axis of rotation also rotates about the vertical direction. This rotation of the top's spin axis is referred to as *precession*.

- 13 There is no reason I have chosen to first interpolate in the y direction and then interpolate in the x direction. I could have first interpolated in the x direction between c_1 and c_2 and then between c_3 and c_4 .

REFERENCES

- Abdelrahman, E.M., Abo-Ezz, E.R. and Essa, K.S., 2012. Parametric inversion of residual magnetic anomalies due to simple geometric bodies. *Exploration Geophysics*, vol. 43, pp. 178–189.
- Abdelrahman, E.M., Soliman, K.S., El-Araby, T.M., Abo-Ezz, E.R. and Essa, K.S., 2009. A least-squares standard deviation method to interpret magnetic anomalies due to thin dykes. *Near Surface Geophysics*, vol. 7, pp. 41–46.
- Aitken, M., Webster, G. and Rees, A., 1958. Magnetic prospecting. *Antiquity*, vol. 32, pp. 270–271.
- Al-Zoubi, A., Haddadin, G. and Batayneh, A., 1998. Geophysical methods for locating a prehistoric site in the Al Hummaymeh area, southwest Jordan. *Journal of Environmental and Engineering Geophysics*, vol. 3, pp. 157–161.
- Anderson, N., Kansas, N.C., Dawood, R.H. and Sirles, P., 2008. Geophysical methods commonly employed for geotechnical site characterization. Transportation Research Board, Transportation research circular E-C130, ISSN 0097-8515.
- Batayneh, A., 2011. Archaeogeophysics—archaeological prospection—a mini review. *Journal of King Saud University—Science*, vol. 23, no. 1, p. 83089.
- Batayneh, A., Al-Zoubi, A., Tobasi, U. and Haddadin, G., 2001. Evaluation of archaeological site potential on the Tall al-Kharrar area (Jordan) using magnetic and electrical resistivity methods. *Environmental Geology*, vol. 41, pp. 54–61.
- Batayneh, A., Khataibeh, J., AlRshdan, H., Tobasi, U. and Al-Jahed, N., 2006. The use of microgravity, magnetometry and resistivity surveys for the characterization and preservation of an archaeological site of Umm er-Rasas. *Jordan Archaeological Prospection*, vol. 14, pp. 60–70.
- Belshe, J., 1957. Recent magnetic investigations at Cambridge University. *Advances in Physics*, vol. 6, pp. 192–193.
- Biswas, A., 2016. Interpretation of gravity and magnetic anomaly over thin sheet-type structure using very fast simulated annealing global optimization technique. *Modeling Earth Systems and Environment*, vol. 2, p. 30.
- Brown, William Fuller, Jr., 1962. *Magnetostatic Principles in Ferromagnetism*. Amsterdam and New York: North-Holland Pub. Co., Interscience Publishers.
- Chapman, S. and Bartels, J., 1940. Geomagnetic and related phenomena. *Geomagnetism*, vol. 1.
- Colani, C. and Aitken, M., 1966. Utilization of magnetic viscosity effects in soils for archaeological prospection. *Nature*, vol. 212, pp. 1446–1447.
- Das, L.K., Das, B., Choudhury, S.N., Naskar, D.C., Karunakar, G. and Dey, S.K., 2005. 3-D configuration of Kimberlite bodies, Indravati Basinal Area, Bastar district, Chhattisgarh. *Geological Society of India*, vol. 65, pp. 679–688.
- Doyle, H.A., Glover, J.E. and Groves D.I. (eds), 1981. *Geophysical Prospecting in Deeply Weathered Terrains*. Geology Department & University Extension, The University of Western Australia, Publication no. 6, p. 179.
- Doyle, H.A. and Lindeman, F.W., 1985. The effect of deep weathering on geophysical exploration in Australia—a review. *Australian Journal of Earth Sciences*, vol. 32, pp. 125–135.
- Essa, K.S., 2011. A new algorithm for gravity or self-potential data interpretation. *Journal of Geophysics and Engineering*, vol. 8, pp. 434–446.
- Essa, K.S. and Munsch, M., 2019. Introductory Chapter: Mineral Exploration from the Point of View of Geophysicists. *Minerals* (Book), DOI: 10.5772/intechopen.84830.
- Essa, K.S., Nady, A.G., Mostafa, M.S. and Elhussein, M., 2018. Implementation of potential field data to depict the structural lineaments of the Sinai Peninsula, Egypt. *Journal of African Earth Sciences*, vol. 147, pp. 43–53.
- Frenez, D., 2014. The Lothal Revisitation Project, a fine thread connecting ancient India to contemporary Ravenna (Via Oman). *BAR Int. Series*, 2690, pp. 263–277.
- Geomagnetic Indices Bulletin, 1990. National geophysical data center. Solar Terrestrial Physics Division (E/GC2). 325 Broadway Boulder, Colorado 80303 USA. S10 1 GE679 V. 1990–1993.

- Gibson, T., 1986. Magnetic prospection on prehistoric sites in western Canada. *Geophysics*, vol. 51, pp. 553–560.
- Grant, I.S.W.R. and Phillips, W.R., 2008. *Electromagnetism*. 2nd edn. Manchester Physics. Hoboken, NJ: John Wiley & Sons, ISBN 9780471927129.
- Jackson, A., Jonkers, A.R.T. and Walker, R.M., 2000. Geomagnetic polarity reversals and long-term secular variation. *Philosophical Transactions: Mathematical, Physical and Engineering Sciences, Article: Four Centuries of Geomagnetic Secular Variation from Historical Records*, vol. 358, no. 1768, pp. 957–990. Published by: Royal Society. www.jstor.org/stable/2666741.
- Kainz, J., 2016. An integrated archaeological prospection and excavation approach at a Middle Neolithic circular ditch enclosure in Austria. In: Forte, M. and Campana, S. (eds), *Digital Methods and Remote Sensing in Archaeology*. Cham, Switzerland: Springer, pp. 371–403.
- Kainz, J., 2019. Addressing archaeological research questions using geophysical surveys—a landscape case study. Research Gate, 13th International Conference on Archaeological Prospection, Sligo, Ireland. Archaeopress Publishing Ltd., DOI: 10.32028/9781789693072.
- Kainz, J. and Cotter, J.P., 2018. A broad band magnetic susceptibility test study—the magnetic spectroscopy of a Neolithic ditch. *Journal of Archaeological Science*, vol. 18, pp. 139–150.
- Kamar, S.A., 2013. *Geophysical Surveying Using Magnetism Methods*. EBS 309. Geofizik Carigali-Magnetik, p. 35. Available at www.earthsci.unimelb.edu.au/ES304/MODULES/MAG/main.html. www.bgu.ac.il/geol/classes/geophysics/Front/Lec12.pdf.
- Kearey, P., Brooks, M. and Hill, I., 2002. *An Introduction to Geophysical Exploration*. 3rd edn. Cambridge: Cambridge University Press; Blackwell Science, ISBN 0632049294.
- Le Maire, P. and Munsch, M., 2018. 2D potential theory using complex algebra: New equations and visualization for the interpretation of potential field data. *Geophysics*, vol. 83, pp. J1–J13.
- Louis, M.N., 1948. “Propriétés magnétiques des ferrites; Ferrimagnétisme et antiferromagnétisme”. *Annales de Physique*, vol. 12, no. 3, pp. 137–198. DOI: 10.1051/anphys/194812030137.
- Majumdar, N., Kumar, N. and Kumar, A., 2016. Geophysical mapping in Nuapada, Bolangir and Kalahandi districts, Odisha. GSI Unpub. Rep. FSP 2015–2016. Toposheet nos. 64 L/11, 12, 14, 15 and 16.
- McGrath, P.H. and Hood, P.J., 1970. The dipping dyke case: A computer curve-matching method of magnetic interpretation. *Geophysics*, vol. 35, pp. 831–848.
- Naskar, D.C., Chakraborty, P., Ramasatish, T. and Tripathi, A., 2020a. Crustal structure from gravity and magnetic Data in parts of Nalgonda and Khammam Districts, Telangana (India). *Journal Indian Geophysical Union (IGU)*, vol. 24, no. 2, pp. 1–6.
- Naskar, D.C., Das, L.K., Majumdar, R.K., Choudhury, K. and Roy, K.K., 2008. Delineation of structural fabric and crustal architectures by gravity interpretation in Central India. *Indian Journal of Geology*, vol. 77, nos. 1–4, pp. 45–53.
- Naskar, D.C., Das, L.K. and Rai, M.K., 2016. Insight into the tectonic and crustal understanding of lesser Himalayas along Purnea-Sevoke transect through geophysical studies. *Journal of Indian Geophysical Union*, vol. 20, no. 5, pp. 506–515.
- Naskar, D.C. and Kulkarni, A., 2019. Geophysical investigation for copper and gold mineralization around Dhani-Basri and Todi ka Bas area, Dausa and Alwar districts, Rajasthan. *Journal of Indian Geophysical Union*, vol. 23, no. 2, pp. 120–133.
- Naskar, D.C., Kulkarni, A. and Lakshmana, M., 2018. Geophysical investigation for base metal mineralization in Karoi-Rajpura area, Bhilwara district, Rajasthan. *Journal of Indian Geophysical Union*, vol. 22, no. 4, pp. 379–388.
- Naskar, D.C. and Lashkari, G., 2019. Geophysical study in Tonk and Dausa districts, Rajasthan: Its implication in regional geology. *Journal Indian Geophysical Union*, vol. 23, no. 6, pp. 485–498.
- Naskar, D.C., Majumdar, N., Ajay, K. and Nikhil, K., 2020b. Geophysical mapping in parts of Nuapada, Bolangir and Kalahandi districts of Odisha (India). *Journal Indian Geophysical Union*, vol. 24, no. 3, pp. 27–41.
- Naskar, D.C., Majumdar, R.K., Choudhury, K. and Das, L.K., 2003. Magnetic profiles across NSL in parts of Central India. *Indian Minerals*, vol. 57, nos. 1–2, pp. 53–60.
- Naskar, D.C. and Rai, M.K., 2018. Delineation of geometry and spatial distribution of basic bodies using magnetic and resistivity methods in Sohagpur coal field area, Madhya Pradesh. *International Journal of Research-Granthaalayah*, vol. 6, no. 8, pp. 146–158, ISSN 2350-0530(O), ISSN 2394-3629(P), DOI: 10.5281/zenodo.1403846.

- Naskar, D.C. and Saha, D.K., 2015. Geophysical investigations for delineation of Gondwana sediments below Deccan trap beyond the Western limit of Wardha Valley Coalfields, Yeotmal and Wardha districts, Maharashtra-a comprehensive analysis of case studies. *Journal of Indian Geophysical Union*, vol. 19, no. 4, pp. 433–446.
- Nettleton, L.L., 1942. Gravity and magnetic calculations. *Geophysics*, vol. 7, pp. 293–310, DOI: 10.1190/1.1445015.
- Prabhakar, V.N. and Korisetar, R., 2017. Ground survey to aerial survey: Methods and best practices in systematic archaeological explorations and excavations. *Current Science*, vol. 113, no. 10, pp. 1873–1890.
- Pridmore, D.F., Coggon, J.H., Esdale, D.J. and Lindemann, F.W., 1984. Geophysical exploration for nickel sulphide deposits in the Yilgarn Block, Western Australia. In: Buchanan, D.L. and Jones, M.J. (eds), *Sulphide Deposits in Mafic and Ultramafic Rocks*. London: Institute of Mining and Metallurgy, pp. 22–34.
- Ravat, D., 2007. Reduction to pole. In: Gubbins, D. and Herrero-Bervera, E. (eds), *Encyclopedia of Geomagnetism and Paleomagnetism*. Dordrecht: Springer.
- Reid, A.B., Allsop, J.M., Granser, H., Millet, A.J. and Somerton, I.W., 1990. Magnetic interpretation in three dimensions using Euler deconvolution. *Geophysics*, vol. 55, pp. 80–91.
- Rivas, J., 2009. Gravity and magnetic methods. Presented at “Short Course on Surface Exploration for Geothermal Resources”, organized by UNU-GTP and LaGeo, in Ahuachapan and Santa Tecla, El Salvador.
- Scollar, I., 1962. Electromagnetic prospecting methods in archaeology. *Archaeometry*, vol. 5, pp. 146–153.
- Somasundaran, P. (ed.), 2006. *Encyclopedia of Surface and Colloid Science*, 2nd edn. New York: Taylor & Francis, p. 3471, ISBN 9780849396083.
- Spaldin, N.A., 2010. 9. Ferrimagnetism. In: *Magnetic Materials: Fundamentals and Applications*. 2nd edn. Cambridge: Cambridge University Press, pp. 113–129, ISBN 9780521886697.
- Stanley, J.M., Sertsrivanit, S. and Clark, P.J., 1992. Magnetic exploration beneath a near-surface magnetic noise source. *Exploration Geophysics*, vol. 23, pp. 323–326.
- Stephen, T., 2011. *Visualization of Fields and Applications in Engineering*. Chichester, West Sussex and Hoboken, NJ: John Wiley and Sons, p. 64, ISBN 9780470978467.
- Tarling, D., Hammo, N. and Downey, W., 1986. The scatter of magnetic directions in Archaeomagnetism studies. *Geophysics*, vol. 51, pp. 634–639.
- Tite, M. and Mullins, C., 1971. Enhancement of the magnetic susceptibility of soils on archaeological sites. *Archaeometry*, vol. 13, pp. 209–219.
- Tiwari, V.M., Veenadhari, B., Gahalaut, V.K., Mukherjee, S. and Dimri, V.P., 2014. Gravity, GPS and geomagnetic data in India. *Proceedings of the Indian National Science Academy*, vol. 80, no. 3, pp. 705–712, DOI: 10.16943/ptinsa/2014/v80i3/55145.
- Trench, A. and Williams, P.K., 2019. NICKEL: Application of geophysics to nickel sulphide exploration in the Kambalda district, Western Australia. *ASEG Extended Abstracts*, *Taylor and Francis*, vol. 1, pp. 169–180.
- VanPool, T.L. and Leonard, R.D., 2011. *Quantitative Analysis in Archaeology*. Hoboken, NJ: John Wiley & Sons.
- Vestine, E.H., 1960. The survey of geomagnetic field in space. *Transactions American Geophysical Union*, vol. 41, DOI: 10.1029/TR041i001p00004.issn:002-8606.
- Weymouth, J., 1986. Archaeological site surveying program at the University of Nebraska. *Geophysics*, vol. 51, pp. 538–552.
- Wolf, A., 1940. Tidal force observations. *Geophysics*, vol. 5, no. 4, pp. 317–402, ISSN (print) 0016-8033, ISSN (online) 1942-2156, PUBLICATION DATA: © 1940 Copyright © 1940 Society of Exploration Geophysicists, DOI: 10.1190/1.1441813.
- Wolfman, D., 1984. Geomagnetic dating methods in archaeology. *Advances in Archaeological Method and Theory*, vol. 7, pp. 363–458.

5 Direct Current (DC) Resistivity Method

In the period from 1912 to 1914 (Dobrin, 1960), Conrad Schlumberger began his pioneering studies which led to an understanding of the merits of utilizing electrical resistivity methods for exploring the subsurface (Compagnie Gbnerale de Gbphysique, 1963). According to Breusse (1963), the real progress in applying electrical methods to groundwater exploration began during World War II. French, Russian, and German geophysicists are mainly responsible for the development of the theory and practice of direct current electrical prospecting methods. Comprehensive overviews of resistivity methods are presented in Telford et al. (1976), Telford et al. (1990), Ward (1990), Kearey and Brooks (1991), Schlumberger (1920), and Burger (1992).

5.1 ELECTRICAL PRINCIPLES

Coulomb established that the force of attraction or repulsion between two charged spheres was proportional to the product of the individual electric charges and inversely proportional to the square of the distance between the centres of the spheres. His law can be written as the following equation:

$$F = k \frac{q_1 q_2}{r^2} \quad (5.1)$$

where q_1 and q_2 are the electric charges, r is their separation, and k is constant. This inverse-square law strongly resembles the law of universal gravitation, formulated by Newton more than a century before Coulomb's law. However, in gravitation, the force is always attractive, whereas in electricity, it may be attractive or repulsive, depending on the nature of the charges.

In Coulomb's law, F and r are defined from mechanics (as the newton and metre, respectively), but the units of q and k are undefined. The value of k was originally set equal to unity, thereby defining the unit of electric charge.

The unit of charge is the *coulomb* (C), defined as the amount of charge that passes a point in an electrical circuit when an electric current of 1 A (ampere) flows for 1 second (i.e. $1C = 1As$). The *ampere* is defined as the current that produces a force of 2×10^{-7} N per metre of length between infinitely long thin conductors that are 1 metre apart in a vacuum. Thus, the unit of charge is defined precisely, if rather indirectly. In the System Internationale (SI) units, k is written as $(\frac{1}{4\pi\epsilon_0})$, so that Coulomb's law becomes

$$F = \frac{1}{4\pi\epsilon_0} \frac{q_1 q_2}{r^2} \quad (5.2)$$

where the constant ϵ_0 is called the 'permittivity constant'. It is approximately equal to $8.854 \times 10^{-12} C^2 N^{(-1)} m^{(-2)}$.

Modern electrical theory descends from the discovery in 1897 by the English physicist Joseph J. Thomson of the electron as the basic elementary unit of electric charge. It has a negative charge of 1.602×10^{-19} C. A proton in the nucleus of an atom has an equal positive charge. Normally, an atom contains as many electrons as it has protons in its nucleus and is electrically neutral. If an atom or molecule loses one or more electrons, it has a net positive charge and is called a positive ion; similarly, a negative ion is an atom or molecule with a surplus of electrons.

In metals, some electrons are only loosely bound to the atoms. They can move with relative ease through a material which is called an electrical conductor. Metals like copper and silver are good conductors. In other materials, called insulators, the electrons are tightly bound to the atoms. Glass, rubber, and dry wood are typical insulators. A perfect insulator does not allow electrons to move through it, whereas a perfect conductor offers no opposition to the passage of electrons. Real conductors offer different degrees of opposition.

A flow of charge, or electric current, results when the free electrons in a conductor move in a common direction. A current of 1 A corresponds to a flow of about 6,250,000,000,000,000 electrons per second past any point of a circuit. The direction of an electric current is defined to be the direction of flow of positive charge, which is opposite to the direction of the motion of the electrons.

5.1.1 ELECTRIC FIELD AND POTENTIAL

The force exerted on a unit electric charge by another charge q is called the electric field of the charge q . Thus, if we let $q_1 = q$ and $q_2 = 1$ in eq. 5.2, we obtain the equation for the electric field E at distance r from a charge q :

$$E = \frac{q}{4\pi\epsilon_0 r^2} \quad (5.3)$$

According to this definition, E has the dimensions of newton/coulomb (NC⁻¹).

The term ‘field’ also has another connotation, introduced by Michael Faraday (1791–1867) to refer to the geometry of the lines of force near a charge. Around a positive point charge, the field lines are directed radially outward, describing the (divergent) direction along which a free positive charge would move (Figure 5.1a); around a negative point charge, they are directed radially inward (convergent) (Figure 5.1b). The field lines of a pair of opposite point charges diverge from the positive charge, spread apart, and converge on the negative charge (Figure 5.1c); they give the appearance of drawing the opposite charges together (Kearey and Brooks, 2002). The combined field of two positive point charges is characterized by field lines that leave each charge and diverge in the space between (Figure 5.1d); the field lines appear visibly to push the like charges apart. The direction of the electric field at any point is tangential to the electric field line. The strength of the field is represented by the spatial concentration of the field lines. Close to either electrical charge, the field is strong, and it weakens with increasing distance from the charge. Consequently, work is required to move a charged particle from one point in the field to another. This work contributes to the potential energy of the system.

For example, at an infinite distance from a positive charge q , the repulsive force on a unit positive charge is zero, but at a distance r , it is given by eq. 5.3. The potential energy of the unit charge at r is called the electric potential at r ; we will denote it U . The units of U are energy per unit charge, that is, joules/coulomb. If we move a distance dr against the field E , the potential changes by an amount dU equal to the work done against E , which is $-E dr$. That is:

$$dU = -E dr \Rightarrow E = - \frac{dU}{dr} \quad (5.4)$$

We can readily compute the electric potential U at r by integration:

$$U = - \int_{\infty}^r E dr = - \int_{\infty}^r \frac{q}{4\pi\epsilon_0 r^2} dr \Rightarrow U = \frac{q}{4\pi\epsilon_0 r} \quad (5.5)$$

The energy needed to move a unit charge from one point to another in the electric field of q is the potential difference between the two points. The unit of potential difference is the same as that of U

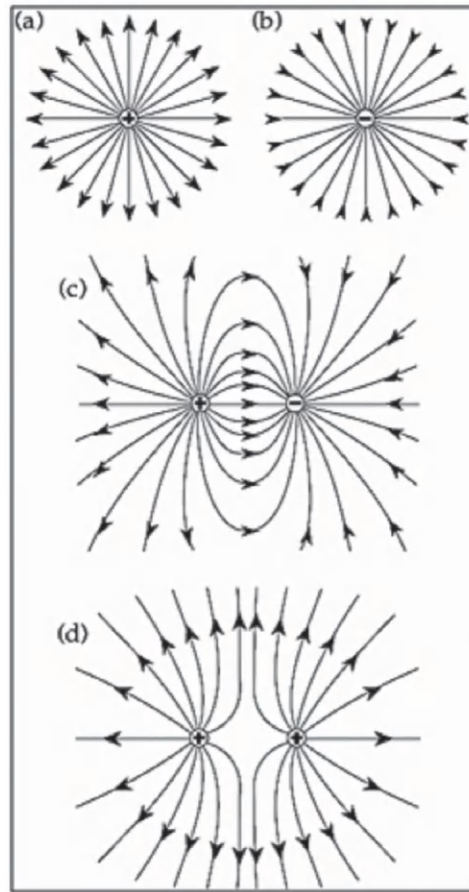


FIGURE 5.1 Planar cross sections of electric field lines around point charges: (a) single positive, (b) single negative, (c) two equal and opposite, and (d) two equal positive charges.

(i.e. joules/coulomb) and is called a volt. From eq. 5.4 we obtain the more common alternative units of volt/metre (Vm^{-1}) for the electric field (E).

An electric charge flows from a point with higher potential to a point with lower potential. The situation is analogous to the flow of water through a pipe from one level to a lower level. The rate of flow of water through the pipe is determined by the difference in gravitational potential between the two levels. Likewise, the electric current in a circuit depends on the potential difference in the circuit.

5.1.2 OHM'S LAW

The German scientist Georg Simon Ohm established in 1827 that the electric current (I) in a conducting wire is proportional to the potential difference (V) across it. The linear relationship is expressed by the equation

$$V = IR \quad (5.6)$$

where R is the resistance of the conductor. The unit of resistance is the ohm (Ω). The inverse of resistance is called the *conductance* of a circuit; its unit is the reciprocal ohm (Ω^{-1}), variously also called a mho or siemens (S).

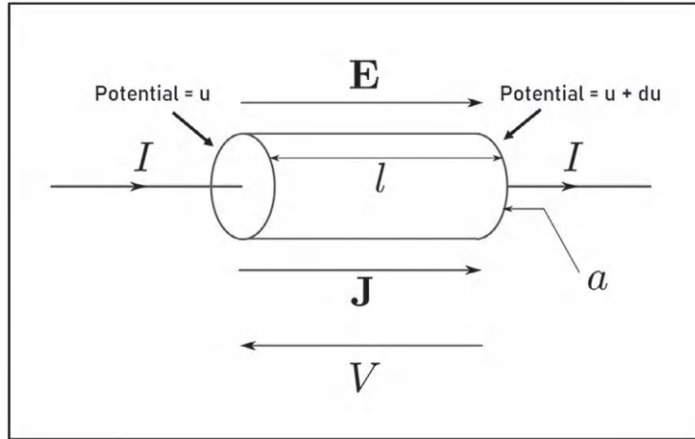


FIGURE 5.2 Parameters used to define Ohm's law for a straight conductor.

Formulated more precisely, for a given material, the resistance is proportional to the length (L) and inversely proportional to the cross-sectional area (A) of the conductor (Figure 5.2). These relationships are expressed in the equation

$$R = \rho \frac{L}{A} \quad (5.7)$$

where ρ , the constant of proportionality, is known as the electrical resistivity or electrical specific resistance, a characteristic of the material which is independent of its shape or size, adapted from Lowrie (2007). It is a physical property of the material of the conductor which expresses its ability to oppose a flow of charge. The inverse of ρ is called the *conductivity* of the material, denoted σ . The unit of resistivity is the ohm-metre (Ωm); the unit of conductivity is the reciprocal ohm-metre ($\Omega^{-1}\text{m}^{-1}$).

If we substitute eq. 5.7 for R in eq. 5.6 and rearrange the terms, we get the following expression:

$$\frac{V}{L} = \rho \frac{I}{A} \quad (5.8)$$

Eq. 5.8 may be used to determine the resistivity ρ of homogeneous and isotropic materials in the form of regular geometric shapes, such as cylinders, parallelepipeds, and cubes. In a semi-infinite material, the resistivity at every point must be defined.

The ratio $\frac{V}{L}$ on the left side of this equation is, by comparison with eq. 5.4, the electric field E (assuming the potential gradient to be constant along the length of the conductor). The ratio $\frac{I}{A}$ is the current per unit cross-sectional area of the conductor; it is called the current density and denoted J (Figure 5.2). We can now rewrite Ohm's law as

$$E = \rho J \quad (5.9)$$

This form is useful for calculating the formulas used in resistivity methods of electrical surveying. However, the quantities that are measured are V and I .

5.1.3 EQUIPOTENTIALS

To describe the nature of electrical current flow in media occupying a volume, we must move beyond our simple notions of current and voltage gained from our experience with wires, resistors, and batteries. In the Earth, or in any three-dimensional body, electrical current is not constrained to flow along a single path as it does in a wire (Jones, 2007). A battery is connected to the Earth by wires and electrodes at two remote points (that is, the electrical connections to the Earth are very distant from one another). The Earth, not being a perfect insulator, conducts the electrical current imparted by the battery. At this stage, let us assume the resistivity of the Earth is uniform throughout the Earth. How does the current flow through the Earth? In this example, current flows (the red lines) out from the electrode (the green rectangle) radially along straight lines (the second electrode is far to the right of the figure as it is drawn). If we could take a voltmeter and measure the voltage drop imposed by the resistivity of the medium from a distance very far from the current electrode to various places in the media, we would find that the voltage drops would be constant along circular lines centred on the electrode (that is, one of the leads to the voltmeter would have to make contact with the ground at a distance very far from the electrode; the other is then moved throughout the medium). These lines are referred to as ‘equipotentials’ (think equal voltage). In three dimensions, they form hemispheres centred on the electrodes. Several equipotential lines are shown in black, with the voltage drop associated with each line shown in greyscale. The darker the greyscale, the smaller the potential drop between this location and a location very far from the current electrode.

5.1.4 ATOMIC CHARGE

Recall that *matter* is conceptualized as being made of atoms: positive-charged nucleus (protons + neutrons) and negative-charged electrons that circle the nucleus in a cloud pattern. Usually, these charges are balanced (e.g. H_2O , NaCl , $\text{KAl}_2\text{Si}_2\text{O}_8$, $[\text{Mg}, \text{Fe}]_2\text{SiO}_4$). An imbalance in charge (i.e. ions) gives a body a net charge (SO_4^{2-} , O^{2-}). *Resistivity* is concerned with the ‘flow’ of charge, not the net charge or any imbalance in charges.

5.1.5 TYPES OF ELECTRICAL CONDUCTION

Electric current passes through a material (conduction) by one of three different modes: by electronic, dielectric, or electrolytic conduction. Electronic (or ohmic) conduction occurs in metals and crystals, dielectric conduction in insulators which contain no free electrons, and electrolytic conduction in liquids.

In rocks (conduction in the Earth), two basic types of conduction occur:

1. *Electronic*. Electrons are mobile in metallic ores and flow freely, for example, metals (wires) and some ore bodies.
2. *Electrolytic/ionic*. Salts disassociate into ions in solution and move, for example, involves motion of cations (+) and anions (−) in opposite directions (Figure 5.3).

5.1.5.1 Dielectric Constant

Normally, the electrons are distributed symmetrically about a nucleus. However, an electric field displaces the electrons in the direction opposite to that of the field, while the heavy nucleus shifts slightly in the direction of the field. The atom or ion acquires an electric polarization and acts like an electric dipole. The net effect is to change the permittivity of the material from ϵ_0 to a different value ϵ , given by

$$\epsilon = \epsilon_r \epsilon_0 \quad (5.10)$$

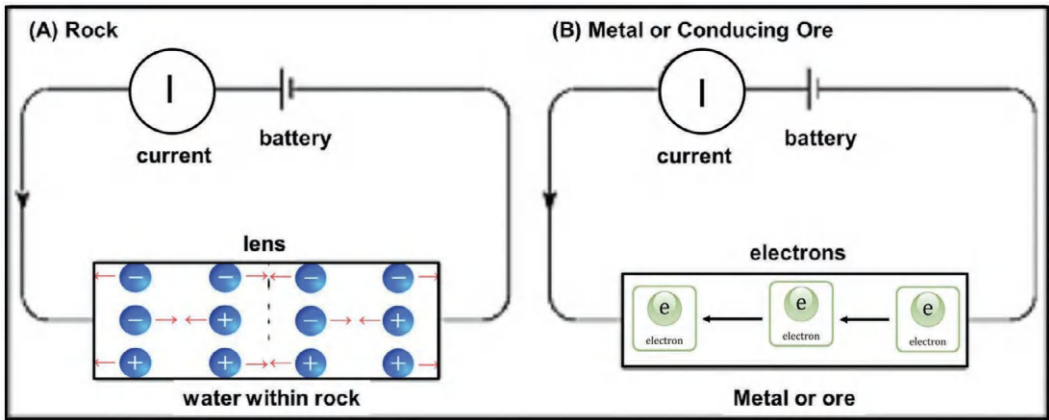


FIGURE 5.3 Conduction process within the Earth.

Source: National Research Council (1996).

Here, ϵ_r is called the ‘relative permittivity’. When it is measured in a constant electric field, the relative permittivity is called the ‘dielectric constant’, k , of the material. It is dimensionless and has a value commonly in the range 3–80. Examples of k for some natural materials are (Table 5.1):

TABLE 5.1	
Dielectric Constant of Natural Materials	
Rock Type	Dielectric Constant (k)
Air	1.00059
Mica	3
Glass	5
Sandstone	5–12
Granite	3–19
Diorite	6
Basalt	12
Water	80

Dielectric effects are unimportant in constant current situations. However, in an alternating electric field, the polarization changes with the frequency of the field, and thus, the relative permittivity is frequency dependent. The fluctuating polarization of the electric charge contributes to the alternating current and so modifies the effective conductivity or resistivity. In practice, this effect depends strongly on the frequency of the inducing alternating field. The higher the frequency, the greater is the effect of dielectric conduction. Some geoelectric methods utilize signals in the audio frequency range, where dielectric conduction is insignificant, but ground-penetrating radar uses frequencies in the MHz to GHz range and depends on dielectric contrasts.

5.1.6 POTENTIALS IN HOMOGENEOUS MEDIA

Consider a continuous current flowing in an isotropic homogeneous medium. (This analysis will also apply to AC if the frequency is low enough that displacement currents are insignificant.) If δA is an element of surface and J the current density in amperes per square metres, then the current

passing through δA is $J \cdot \delta A$. The current density (J) and the electric field (E) are related through Ohm's law:

$$J = \sigma E$$

where E is in volts per metre and σ is the conductivity of the medium in siemens per metre (S/m).

The electric field is the gradient of a scalar potential:

$$E = -\nabla V$$

Thus, we have $J = -\sigma \nabla V$.

The original Ampere's law is simply mathematically inconsistent, unless the electric current has zero divergence, $\nabla \cdot J = 0$. So we have

$$\nabla \cdot \sigma \nabla V = 0 \Rightarrow \nabla \sigma \cdot \nabla V + \sigma \nabla^2 V = 0$$

If σ is constant throughout, the first term vanishes and we have Laplace's equation, that is, the potential is harmonic:

$$\nabla^2 V = 0$$

There are two boundary conditions that must hold at any contact between two regions of different conductivity, where σ and μ change abruptly:

$$E_{x_1} = E_{x_2} \text{ and } \sigma_1 E_{z_1} = \sigma_2 E_{z_2} \quad (5.11a)$$

where the x and z axes are tangential and normal, respectively, to the interface, E_{x_1} , being the tangential component in medium 1, and so forth. In addition:

$$V_1 = V_2 \quad (5.11b)$$

5.1.7 SINGLE-CURRENT ELECTRODE AT DEPTH

There are several field configurations used in resistivity that we will consider in turn. In the first of these we have an electrode of small dimensions buried in a homogeneous isotropic medium. This corresponds to the mise-a-la-masse method, where the single electrode is down a drill hole or otherwise under the ground. The current circuit is completed through another electrode, usually at the surface, but in any case far enough away that its influence is negligible.

From the symmetry of the system, the potential will be a function of r only, where r is the distance from the first electrode. Under these conditions, Laplace's equation, in spherical coordinates, simplifies to

$$\nabla^2 V = \frac{d^2 V}{dr^2} + \frac{2}{r} \frac{dV}{dr} = 0$$

Multiplying by r^2 and integrating, we get

$$\frac{dV}{dr} = \frac{A}{r^2}$$

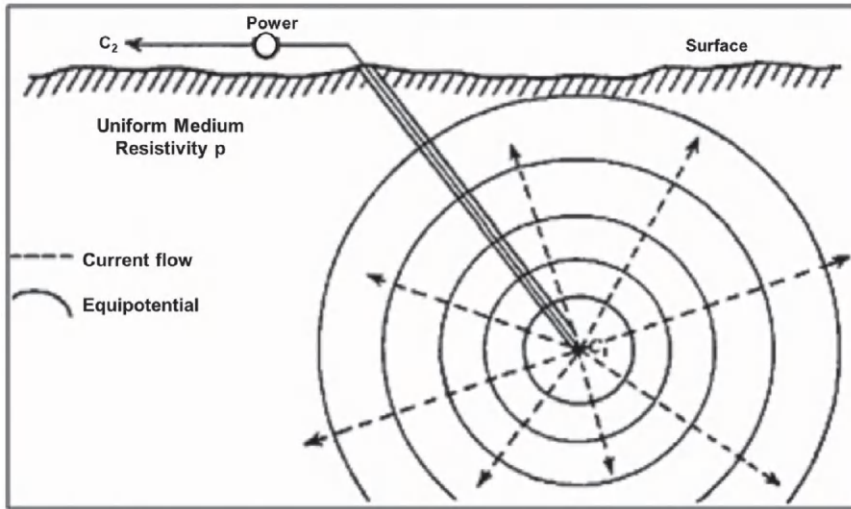


FIGURE 5.4 Buried point source of current in homogeneous ground.

Integrating again, we have

$$V = -\frac{A}{r} + B$$

where A and B are constants. Because $V = 0$ when $r \rightarrow \infty$, we get $B = 0$. In addition, the current flows radially outward in all directions from the point electrode. Thus, the total current crossing a spherical surface is given by

$$I = 4\pi r^2 J = -4\pi r^2 \sigma \frac{dV}{dr} = -4\pi \sigma A \Rightarrow A = -\frac{I}{4\pi \sigma} = -\frac{I\rho}{4\pi}$$

Hence, by substitution we get

$$V = \frac{I\rho}{4\pi r} \Rightarrow \rho = \frac{4\pi r V}{I} \quad (5.12)$$

The equipotentials, which are everywhere orthogonal to the current flow lines, will be spherical surfaces given by $r = \text{constant}$. These are illustrated in Figure 5.4 (Nimmer and Osiensky, 2002; Rahmani et al., 2020).

5.2 ARCHIE'S LAW

Porous, water-bearing rocks/sediments may be ionic conductors. Their 'formation resistivity' is defined by Archie's law:

$$\rho_t = a \cdot \rho_w \varphi^{-m} s_w^{-n}$$

where φ = porosity, s_w = water saturation, $a \approx 0.5$ – 2.5 , $n \approx 2$ if $s_w \geq 3$, m = cementation ≈ 1.3 (Tertiary) and 2 for Palaeozoic. Archie's law is an empirical model. Exponents imply about the range of resistivity of geologic materials (Archie, 1942). The following table (Table 5.2) shows the porosity of rock (Keller, 1966).

TABLE 5.2
Porosity of Rocks

Rock Type	Usual Range of Porosity (%)	
	From	To
Shale	20	50
Sand	20	35
Sandstone	5	30
Limestone	1.5	15
Dolomite	3	20
Quartzite		0.8
Granite		1.2
Diabase		1
Biotite granite	0.4	5.2
	0.7	5
Nepheline syenite	1.3	5.1
Diorite	1.3	5.1
Gabbro	0.3	3.5
Norite gabbro	1.3	2

5.3 ROCK RESISTIVITIES

The important physical properties of rocks for electrical surveying are permittivity (for georadar) and resistivity (or conductivity), on which several techniques are based.

The resistivity ρ of rocks and minerals displays a wide range. For example, native silver has a resistivity of the order of 10^{-8} Ω -m (least resistive), graphite has a resistivity of the order of 10^{-6} Ω -m, some dry quartzite rocks have resistivities of more than 10^{12} Ω -m, whereas pure sulphur has a resistivity of the order of 10^{16} Ω -m (most resistive) (Parasnis, 1962). No other physical property of naturally occurring rocks or soils displays such a wide range of values.

In most rocks, electricity is conducted electrolytically by the interstitial fluid, and resistivity is controlled more by porosity, water content, and water quality than by the resistivities of the rock matrix. Clay minerals, however, are capable of conducting electricity electronically, and the flow of current in a clay layer is both electronic and electrolytic. Resistivity values for unconsolidated sediments commonly range from less than 1 Ω -m for certain clays or sands saturated with saline water to several thousand ohm-metre for dry basalt flows, dry sand, and gravel. The resistivity of sand and gravel saturated with fresh water ranges from about 15 to 600 Ω -m.

The values ranging from 15 to 20 Ω -m are characteristic of aquifers in the South-Western United States, whereas in certain areas in California, the resistivity of freshwater-bearing sands generally ranges from 100 to 250 Ω -m. In parts of Maryland, resistivities have been found to range between about 300 and 600 Ω -m, which is about the same range as that for basaltic aquifers in southern Idaho (Zohdy et al., 1969). The following table shows the resistivities of rock and minerals (Table 5.3).

Values or ranges, which have come from several sources, are only approximate, with older rocks having the highest resistivity due to more time to fill in fractures and pore space, while younger rocks have the lowest resistivity due to abundant fractures and/or pore space. The resistivity contrast between the ore body and the host rock is often large, because the resistivities of different rocks and minerals vary widely (Figure 5.5).

TABLE 5.3**Resistivities of Some Rocks and Minerals**

Rocks, Minerals, Ores	Resistivity (Ohm-Metre)
Sediments	
Chalk	50–150
Clay	1–100
Gravel	100–5,000
Limestone	50–10 ⁷
Marl	1–100
Quartzite	10–10 ⁸
Shale	10–1,000
Sand	500–5,000
Sandstone	1–10 ⁸
Igneous and Metamorphic Rocks	
Basalt	10–10 ⁷
Gabbro	1,000–10 ⁶
Granite	100–10 ⁵
Marble	100–10 ⁸
Schist	10–10 ⁴
Slate	100–10 ⁷
Minerals and Ores	
Silver	1.6–10 ⁻⁸
Graphite, massive ore	10 ⁻⁴ –10 ⁻³
Galena (PbS)	10 ⁻³ –10 ²
Magnetite ore	1–10 ⁵
Sphalerite (ZnS)	10 ³ –10 ⁶
Pyrite	1 × 100
Chalcopyrite	1 × 10 ⁻⁵ –0.3
Quartz	10 ¹⁰ –2 × 10 ¹⁴
Rock salt	10–10 ¹³
Waters and Effect of Water and Salt Content	
Pure water	1 × 10 ⁶
Natural water	1–10 ³
Seawater	0.2
20% salt	5 × 10 ⁻²
Granite, 0% water	10 ¹⁰
Granite, 0.19% water	1 × 10 ⁶
Granite, 0.31% water	4 × 10 ³

5.3.1 GENERAL RULES OF THUMB FOR RESISTIVITY

Highest resistivity: igneous rocks (only a minor component of pore water) and older rocks (more time to fill in fractures and pore space)

Intermediate resistivity: metamorphic rocks (hydrous minerals and fabrics)

Lowest resistivity: sedimentary rocks (abundant pore space and fluids and clay exhibit super low resistivity) and younger rocks (abundant fractures and/or pore space)

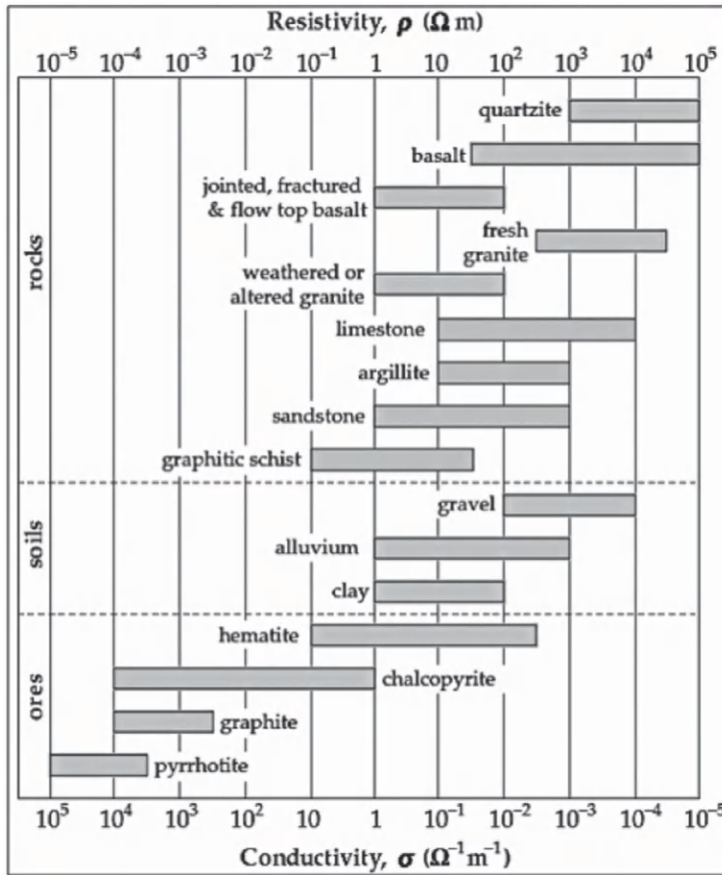


FIGURE 5.5 Ranges of electrical resistivity for some common rocks, soils, and ores.

Source: Ward (1990); augmented by data from Telford et al. (1990); Choudhury et al. (1997).

5.4 CURRENT DISTRIBUTION

The current pattern in a uniform half-space extends laterally on either side of the profile line. Viewed from above, the current lines bulge outward between source and sink with a geometry similar to that shown in Figure 5.1c. In a vertical section, the current lines resemble half of the dipole geometry. In three dimensions, the current can be visualized as flowing through tubes that flatten as they leave the source and narrow as they converge towards the sink. Figure 5.6 shows the flow pattern of the current in a vertical section through the ‘tubes’ in a uniform half-space.

In order to evaluate the depth penetration of the current in a uniform half-space, we define orthogonal Cartesian coordinates with the x axis parallel to the profile and the z axis vertical (Figure 5.7a). Let the spacing of the current electrodes be L , and the resistivity of the half-space be ρ (Telford et al., 1990). The horizontal electric field E_x at (x, y, z) is

$$E_x = -\frac{\partial U}{\partial x} = -\frac{\partial}{\partial x} \left\{ \frac{\rho I}{2\pi} \left(\frac{1}{r_1} - \frac{1}{r_2} \right) \right\}$$

where $r_1 = \sqrt{x^2 + y^2 + z^2}$ and $r_2 = \sqrt{(L-x)^2 + y^2 + z^2}$.

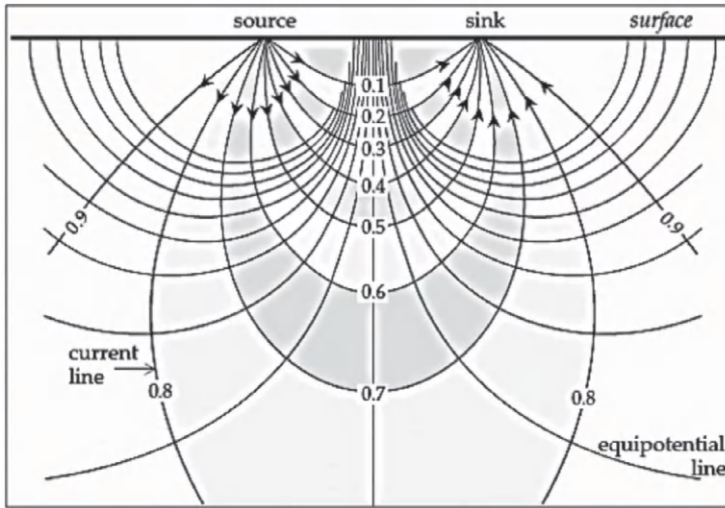


FIGURE 5.6 Cross section of current ‘tubes’ and equipotential surfaces between a source and sink. Numbers on the current lines indicate the fraction of current flowing above the line.

Source: Robinson and Coruh (1988), based upon Van Nostrand and Cook (1966).

Differentiating and using Ohm’s law (eq. 5.9) gives the horizontal current density J_x at (x, y, z) :

$$J_x = \frac{I}{2\pi} \left(\frac{x}{r_1^3} + \frac{L-x}{r_2^3} \right)$$

If (x, y, z) is on the vertical plane midway between the current electrodes, $x = \frac{L}{2}$, $r_1 = r_2$, and the current density is given by

$$J_x = \frac{IL}{2\rho} \frac{1}{\left\{ \left(\frac{L}{2} \right)^2 + y^2 + z^2 \right\}^{\frac{3}{2}}}$$

The horizontal current dI_x across an element of area $(dydz)$ in the median vertical plane is $dI_x = J_x dydz$. The fraction of the input current I that flows across the median plane above a depth z is obtained by integration:

$$\frac{I_x}{I} = \frac{L}{2\pi} \int_0^z dz \int_{-\infty}^{+\infty} \frac{dy}{\left\{ \left(\frac{L}{2} \right)^2 + y^2 + z^2 \right\}^{\frac{3}{2}}} = \frac{L}{\pi} \int_0^z \frac{dz}{\left(\frac{L}{2} \right)^2 + z^2} = \frac{2}{\pi} \tan^{-1} \frac{2z}{L}$$

Eq. 5.18 shows that I_x depends upon the current electrode spacing L (Figure 5.7b). Half the current crosses the plane above a depth $z = \frac{L}{2}$, and almost 90% passes above the depth $z = 3L$. The fraction of current between any two depths is found from the difference in the fractions above each depth calculated with eq. 5.18.

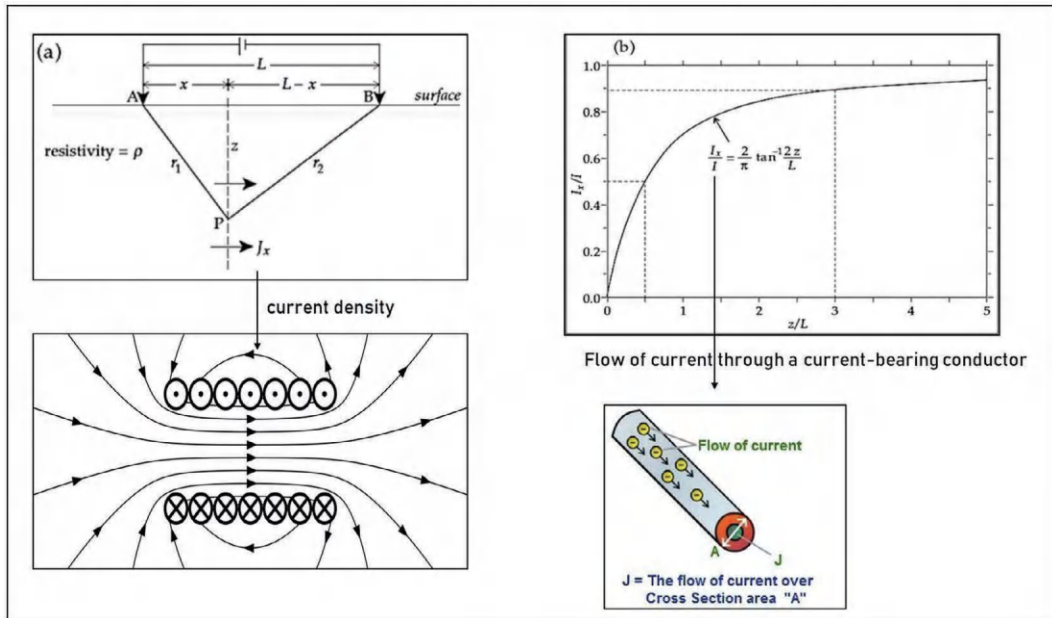


FIGURE 5.7 (a) Geometry for determining the current density in uniform ground below two electrodes, and (b) fraction of current $\left(\frac{I_x}{I}\right)$ that flows above depth z across the median plane between current electrodes with spacing L .

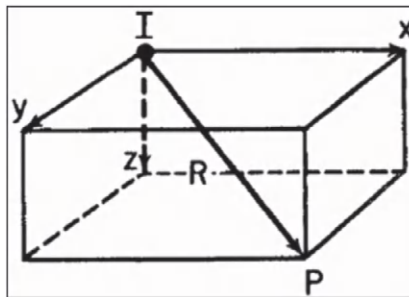


FIGURE 5.8 Diagram showing the relationship between a point source of current I (at origin of coordinates) in an isotropic medium of resistivity ρ and the potential V at any point P .

Source: Kearey and Brooks (2002).

5.5 PRINCIPLES OF RESISTIVITY METHOD

In making resistivity surveys, a commutated direct current or very-low-frequency (<1 Hz) current is introduced into the ground via two electrodes. The potential difference is measured between a second pair of electrodes. If the four electrodes are arranged in any of several possible patterns, the current and potential measurements may be used to calculate resistivity. The electric potential V at any point P caused by a point electrode emitting an electric current I (Figure 5.8) in an infinite homogeneous and isotropic medium of resistivity ρ is given by (from eq. 5.12)

$$V = \frac{\rho I}{4\pi R} \text{ where, } R = \sqrt{x^2 + y^2 + z^2} \quad (5.13)$$

For a semi-infinite medium (Figure 5.9) (Kearey and Brooks, 2002), which is the simplest Earth model, and with both current and potential point electrodes placed at the Earth surface ($z = 0$), eq. 5.13 reduces to

$$V = \frac{\rho I}{2\pi\sqrt{x^2 + y^2}} = \frac{\rho I}{2\pi AM} \quad (5.14)$$

where AM is the distance on the Earth surface between the positive current electrode A and the potential electrode M .

When two current electrodes, A and B , are used and the potential difference (ΔV) is measured between two measuring electrodes M and N (Figure 5.10), we get:

$$V_M^A = \frac{\rho I}{2\pi} \cdot \frac{1}{AM} = \text{potential at } M \text{ due to positive electrode } A$$

$$V_N^A = \frac{\rho I}{2\pi} \cdot \frac{1}{AN} = \text{potential at } N \text{ due to positive electrode } A$$

$$V_M^B = \frac{\rho I}{2\pi} \cdot \frac{1}{BM} = \text{potential at } M \text{ due to negative electrode } B$$

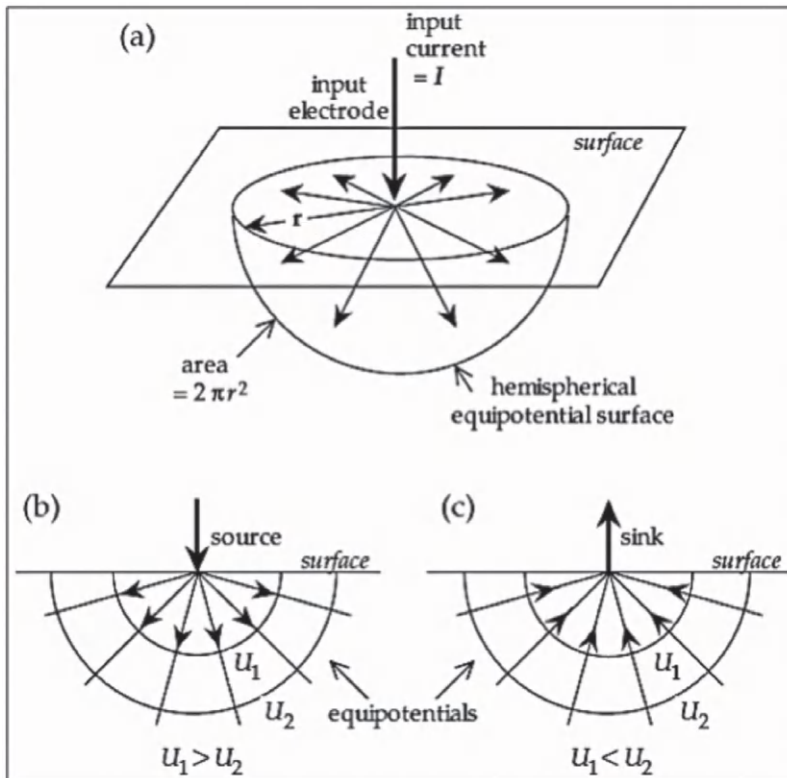


FIGURE 5.9 Electric field lines and equipotential surfaces around a single electrode at the surface of a uniform half-space: (a) hemispherical equipotential surfaces, (b) radially outward field lines around a source, and (c) radially inward field lines around a sink.

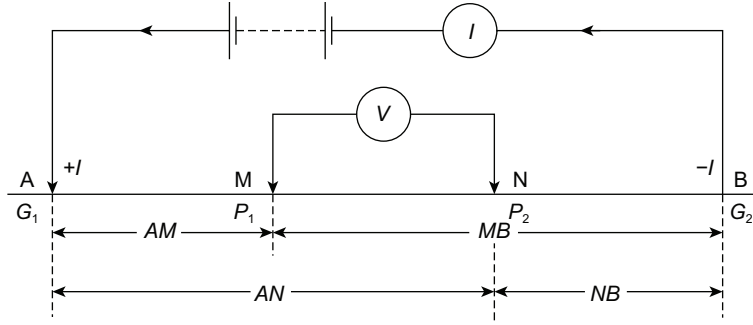


FIGURE 5.10 The general case is considered, where the current sink is a finite distance from the source.

$$V_N^B = \frac{\rho I}{2\pi} \cdot \frac{1}{BN} = \text{potential at } M \text{ due to negative electrode } B$$

$$V_M^{A,B} = \frac{\rho I}{2\pi} \cdot \left(\frac{1}{AM} - \frac{1}{BM} \right) = \text{total potential at } M \text{ due to } A \text{ and } B$$

$$V_N^{A,B} = \frac{\rho I}{2\pi} \cdot \left(\frac{1}{AN} - \frac{1}{BN} \right) = \text{total potential at } N \text{ due to } A \text{ and } B$$

And therefore, the net potential difference is:

$$\Delta V_{MN}^{A,B} = V_M^{A,B} - V_N^{A,B} = \frac{\rho I}{2\pi} \left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN} \right) \quad (5.15)$$

Such an arrangement corresponds to the four electrode spreads normally used in resistivity field-work. In this configuration, the current flow lines and equipotentials are distorted by the proximity of the second-current electrode B . The equipotentials and orthogonal current lines obtained by plotting the relations $\frac{1}{R_1} - \frac{1}{R_2} = \text{constant}$ and $R_1^2 + R_2^2 - 2R_1R_2\cos\theta = 4L^2$ are shown in Figure 5.11 (Telford et al., 1990). The distortion from spherical equipotentials is most evident in the regions between the current electrodes. Rearranging eq. 5.15, we express the resistivity ρ by:

$$\rho = \frac{2\pi}{\left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN} \right)} \frac{\Delta V}{I} \quad (5.16)$$

Eq. 5.16 is a fundamental equation in direct current (DC) electrical prospecting. The factor

$\frac{2\pi}{\left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN} \right)}$ is called the geometric factor of the electrode arrangement and gener-

ally is designated by the letter K . Therefore:

$$\rho = K \frac{\Delta V}{I} \quad (5.17)$$

If the measurement of ρ is made over a semi-infinite space of homogeneous and isotropic material, then the value of ρ computed from eq. 5.16 will be the true resistivity of that material. However,

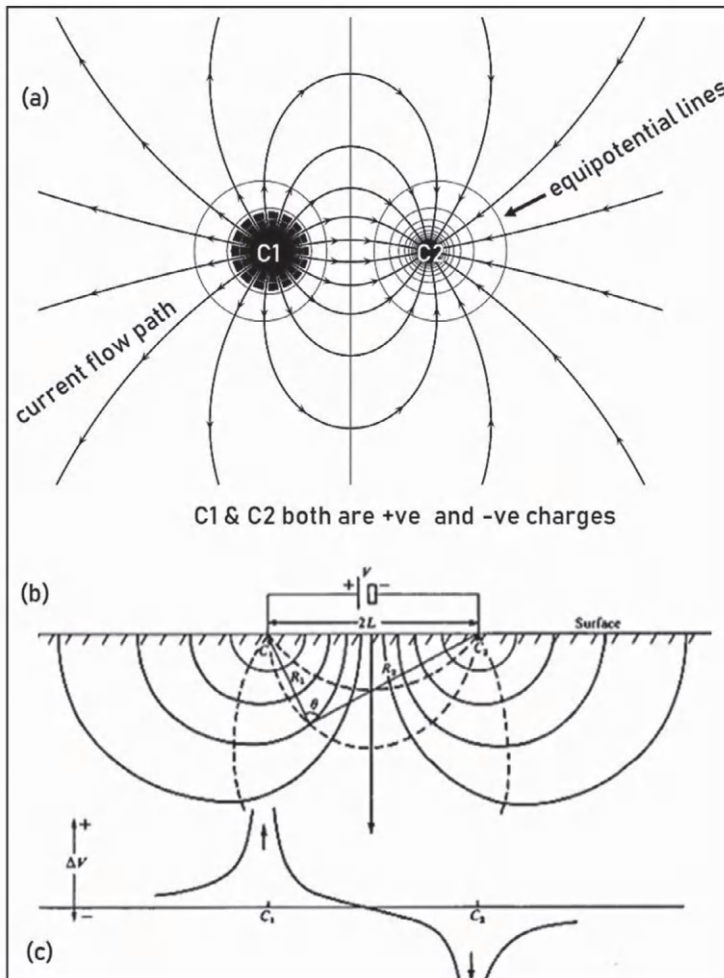


FIGURE 5.11 Equipotentials and current flow lines (paths) for two point sources of current on the surface of a homogeneous ground: (a) plan view, (b) vertical section, (c) potential variation at the surface along a straight line through the point sources.

if the medium is inhomogeneous and/or anisotropic, then the resistivity computed from eq. 5.16 is called an apparent resistivity ρ_a .

The value of the apparent resistivity is a function of several variables: the electrode spacings AM , AN , BM , and BN ; the geometry of the electrode array; and the true resistivities and other characteristics of the subsurface materials, such as layer thicknesses, angles of dip, and anisotropic properties. The apparent resistivity, depending on the electrode configuration and on the geology, may be a crude average of the true resistivities in the section, may be larger or smaller than any of the true resistivities, or may even be negative (Alpin, 1950; Zohdy, 1969b).

Consider a horizontally layered structure in which a layer of thickness d and resistivity ρ_1 overlies a conducting half-space with a lower resistivity ρ_2 (Figure 5.12) (Reynolds, 1998). If the current electrodes are close together, so that $L \ll d$, all or most of the current flows in the more resistive upper layer, so that the measured resistivity is close to the true value of the upper layer ρ_1 . With increasing separation of the current electrodes, the depth reached by the current lines increases. Proportionally, more current flows in the less-resistive layer, so the measured resistivity decreases. Conversely, if the upper layer is a better conductor than the lower layer, the apparent resistivity

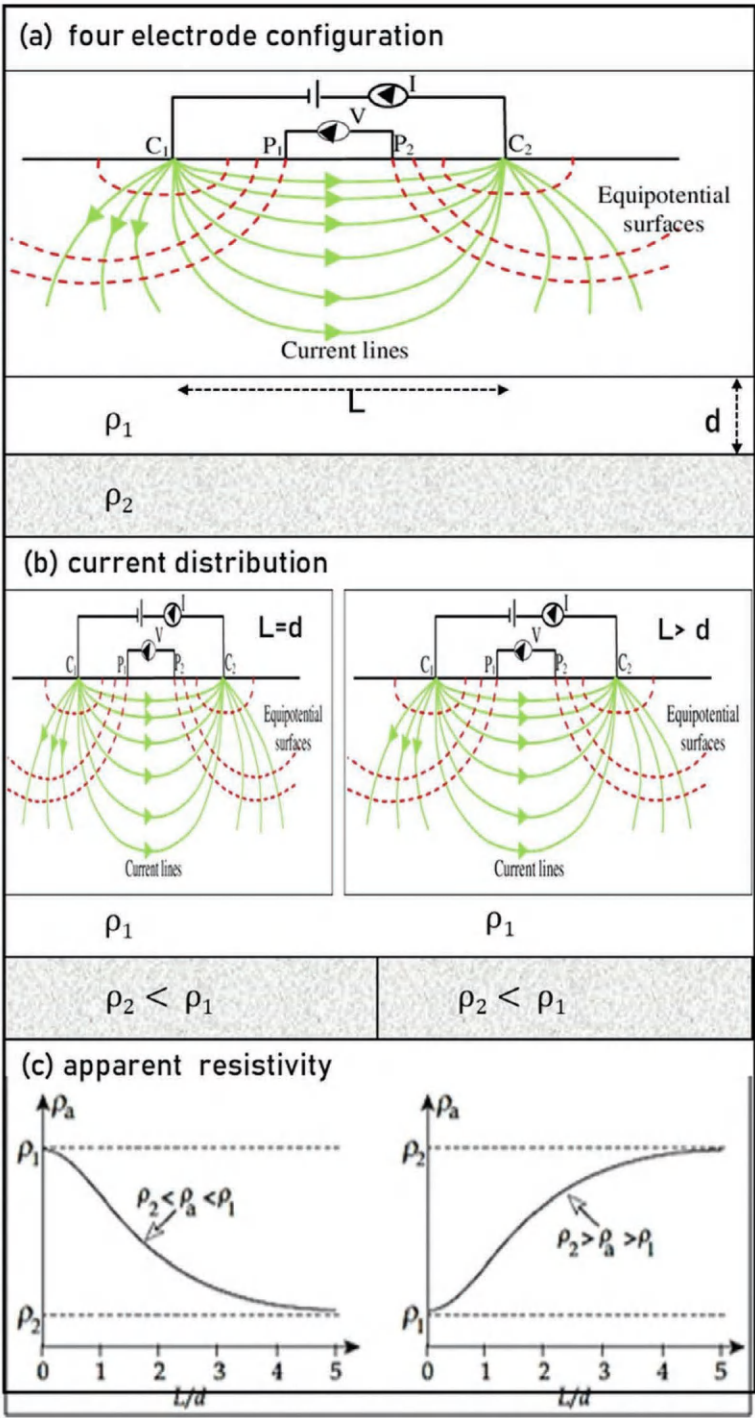


FIGURE 5.12 (a) Parameters of the four-electrode arrangement, (b) distribution of current lines in a two-layer ground with resistivities ρ_1 and ρ_2 ($\rho_1 > \rho_2$), and (c) the variation of apparent resistivity as the current electrode spacing is varied for the two cases of $\rho_1 > \rho_2$ and $\rho_1 < \rho_2$.

Source: Reynolds (1998).

increases with increasing electrode spacing. When the electrode separation is much larger than the thickness of the upper layer ($L \gg d$), the measured resistivity is close to the value ρ_2 of the bottom layer. Between the extreme situations, the apparent resistivity determined from the measured current and voltage is not related simply to the true resistivity of either layer.

5.6 ELECTRODE CONFIGURATIONS

In any case, the geometric factor for any four-electrode system can be found from eq. 5.15 and can be developed for more complicated systems by using the rule illustrated by eq. 5.15. The value of ρ_a (eq. 5.16) depends on the four distance variables AM , AN , BM , and BN . If ρ_a is made to depend on only one distance variable, the number of theoretical curves can be greatly reduced. Several electrode arrays have been invented to fulfil this goal. It can also be seen from eq. 5.17 that the current and potential electrodes can be interchanged without affecting the results; this property is called *reciprocity* (Figure 5.13) (Kearey and Brooks, 2002).

5.6.1 WENNER ARRAY

This well-known array was first proposed for geophysical prospecting by Wenner (1915). The four electrodes A , M , N , and B are placed at the surface of the ground along a straight line (Figure 5.14) so that $AM = MN = NB = a$. For the Wenner array, eq. 5.16 reduces to:

$$\rho_a = 2\pi a \frac{\Delta V}{I} \quad (5.18)$$

Thus, the resistivity ρ_a is a function of the single distance variable, a . The Wenner array is widely used in the Western Hemisphere.

The Wenner electrode array is used for profiling or mapping for delineation of active fault and depth to the bedrock. However, it is possible to get a picture of the subsurface in terms of resistivity for finding the best ground bed location. With today's technology, it would probably be better,

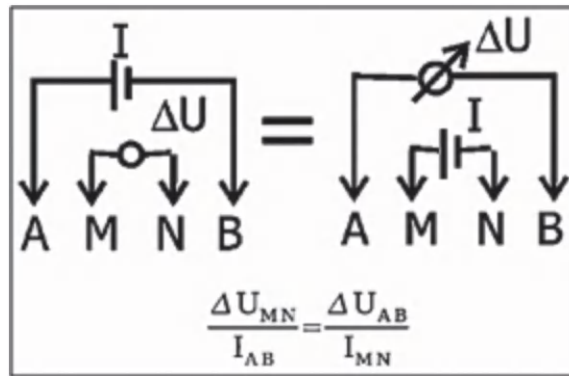


FIGURE 5.13 Property of reciprocity.

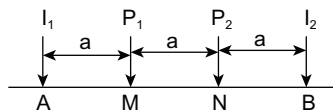


FIGURE 5.14 Wenner electrode array configuration for resistivity measurement.

based on the result and cost, to perform automatic electrical resistivity imaging surveys instead of the standardized resistivity surveys (Naskar et al., 2011).

5.6.2 LEE-PARTITIONED ARRAY

This array is the same as the Wenner array, except that an additional potential electrode O is placed at the centre of the array between the potential electrodes M and N . Measurements of the potential difference are made between O and M and between O and N (Figure 5.15). The formula for computing the Lee-partitioning apparent resistivity is given by

$$\frac{\Delta V}{I} \quad (5.19)$$

where ΔV is the potential difference between O and M or O and N . This array has been used extensively in the past (Van Nostrand and Cook, 1966).

5.6.3 SCHLUMBERGER ARRAY

This array is the most widely used in electrical prospecting. Four electrodes are placed along a straight line on the Earth's surface (Figure 5.16) in the same order, AMNB, as in the Wenner array, but with $MN \leq \frac{1}{5} AB$.

For any linear, symmetric array AMNB of electrodes, eq. 5.16 can be written in the form

$$\rho_a = \pi \frac{\left(\frac{AB}{2}\right)^2 - \left(\frac{MN}{2}\right)^2}{MN} \cdot \frac{\Delta V}{I} \quad (5.20)$$

But if $MN \rightarrow 0$, then eq. 5.20 can be written as

$$\rho_a = \pi \left(\frac{AB}{2}\right)^2 \frac{E}{I} \quad (5.21)$$

where $E = \lim_{MN \rightarrow 0} \frac{\Delta V}{MN}$ = electric field

Conrad Schlumberger defined *resistivity* in terms of the electric field (E) rather than the potential difference (ΔV) (as in the Wenner array). It can be seen from eq. 5.21 that the Schlumberger

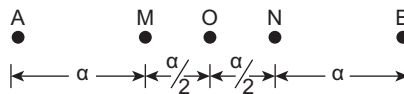


FIGURE 5.15 Lee-partitioning electrode array.

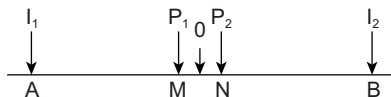


FIGURE 5.16 Schlumberger electrode array configuration for resistivity measurement.

apparent resistivity ρ_a is a function of a single distance variable $\left(\frac{AB}{2}\right)$. In practice, it is possible to measure ρ_a according to eq. 5.21, but only in an approximate manner. The apparent resistivity ρ_a usually is calculated by using eq. 5.20, provided that $MN \leq \frac{1}{5} AB$ (Deppermann, 1954). The Schlumberger array is commonly used for vertical electrical sounding (VES) for groundwater and aggregate minerals. (In mineralogy and petrology, an *aggregate* is a mass of mineral crystals, mineraloids particles, or rock particles. Examples include dolomite, which is an aggregate of crystals of the mineral dolomite, and rock gypsum, an aggregate of crystals of the mineral gypsum.) VES using the Schlumberger array provides better resolutions.

The Schlumberger array is also most commonly used with large spacing at both ends for deeper subsurface information in order to delineate the geoelectrical characteristics of basement complex and sedimentary formation (Zohdy, 1969a, 1969b; Naskar and Rai, 2018; Naskar et al., 2006; Hassan et al., 2017). Schlumberger electrical resistivity sounding is very effective for distinct geophysical anomalies of top layers over a pipe suspected to be kimberlite in the Precambrian metamorphic terrain (Verheijen and Ajakaiye, 1979; Das et al., 2005) after detailed interpretation of the gravity and magnetic data determined the structure to a depth of 0.5 km.

5.6.4 POLE-DIPOLE ARRAY

A *pole* is a single transmitting electrode, and a *dipole* is a pair of oppositely charged electrodes that are so close together that the electric field seems to be a single electrode field instead of fields from two different electric poles (Figure 5.17). For the pole-dipole array, eq. 5.16 reduces to

$$\rho_a = 2\pi a n(n+1) \frac{V}{I} \quad (5.22)$$

For the pole-dipole array with $n = 1$, we get three electrode arrays. The apparent resistivity is given as

$$\rho_a = 4\pi a \left[\frac{V}{I} \right] \quad (5.23)$$

where $C_1 P_1 = P_1 P_2 = a$ and $C_2 = \infty$.

5.6.5 POLE-POLE ARRAY

The four electrodes in the pole-pole array are arranged so that the distance between the transmitter (A and B) dipole and the receiver (M and N) dipole is small in comparison. In a pole-dipole survey,

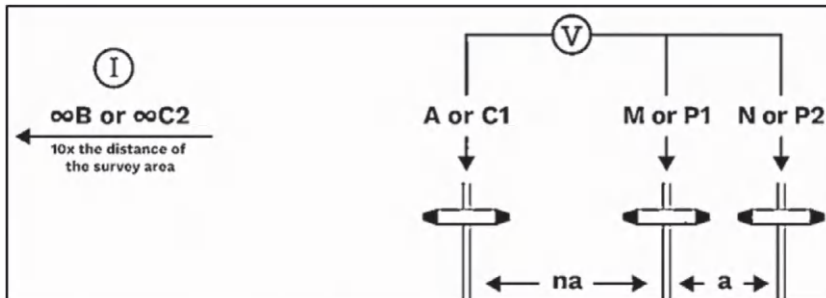


FIGURE 5.17 Pole-dipole electrode array configuration for resistivity measurement.

one transmitting electrode is moved away from the dipole to infinity (i.e. ten times the distance of the survey area) so the instrument does not sense the pole electrode. In a pole–pole survey, one receiver electrode is also moved to infinity, but additionally, one of the potential electrodes is moved to infinity in the opposite direction (Figure 5.18). In other words, in a pole–pole array, we have a stationary infinity electrode on either side of the survey area. For the pole–pole array, eq. 5.16 reduces to

$$\rho_a = 2\pi a \left[\frac{V}{I} \right] \quad (5.24)$$

5.6.6 GRADIENT ARRAY

The gradient array is similar to the Schlumberger array. The difference is that the Schlumberger array records only the centre receiver dipole, whereas the gradient array measures all adjacent dipoles from one transmitter electrode to the other, including the centre-most dipole (Figure 5.19). With the gradient array, the spacing of the outer two electrodes is kept constant, while the two inner electrodes (the potential electrodes) are moved. The spacing between the inner electrodes is constant, but they are moved as a pair in the space between the outer electrodes and measure the potential as they go.

5.6.7 DIPOLE–DIPOLE ARRAYS

The use of dipole–dipole arrays in electrical prospecting has become common since the 1950s, particularly in Russia, where Al'pin (1950) developed the necessary theory. In a dipole–dipole array, the distance between the current electrodes *A* and *B* (current dipole) and the distance between the potential electrodes *M* and *N* (measuring dipole) are significantly smaller than the distance *r* between the centres of the two dipoles. The figure in Naskar and Das (2009) shows the four basic dipole–dipole arrays that are recognized: azimuthal, radial, parallel, and perpendicular. When the azimuth angle θ formed by the line *r* and the current dipole *AB* equals $\frac{\pi}{2}$, the azimuthal array and the parallel array reduce to the equatorial array, and when $\theta = 0$, the parallel and radial arrays reduce to the polar (or axial) array. It can be shown (Al'pin, 1958; Bhattacharya and Patra, 1968; Keller and Frischknecht, 1966) that the electric field, due to a dipole at a given point, is inversely proportional to the cube of the distance *r*, and that for a given azimuth angle θ , the value of the apparent resistivity ρ_a is a function of the single distance variable *r*.

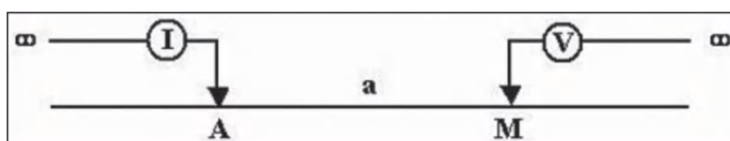


FIGURE 5.18 Pole–pole electrode array configuration for resistivity measurement.

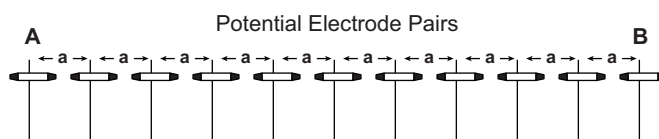


FIGURE 5.19 Gradient electrode array configuration for resistivity measurement.

Of the various dipole–dipole arrays, the equatorial array, in its bipole–dipole form (AB is large, and MN is small), has been used more often than the other dipole–dipole arrays. By enlarging the length of the current dipole, that is, by making it a *bipole*, the electric current required to generate a given potential difference (ΔV) at a given distance r from the centre of the array is reduced. Furthermore, the apparent resistivity remains a function of the single distance variable,

$\bar{R} = \sqrt{\left(\frac{AB}{2}\right)^2 + r^2}$ (Berdichevskii and Petrovskii, 1956). The equatorial array has been used extensively by Russian geophysicists in petroleum exploration (Berdichevskii and Zagarmistr, 1958). It has been used in groundwater investigations in the United States (Zohdy and Jackson, 1969; Zohdy, 1969a).

The approximate geometric factors according to eq. 5.16 for different bipole–dipole configurations are as follows (Alpin et al., 1966):

$$K_{\text{parallel}} = \frac{2\pi r^3}{Ll} \left[\frac{1}{3\cos^2\theta - 1} \right] \quad (5.25)$$

$$K_{\text{perpendicular}} = \frac{2\pi r^3}{3Ll} \left[\frac{1}{\sin\theta\cos\theta} \right] \quad (5.26)$$

$$K_{\text{radial}} = \frac{\pi r^3}{Ll\cos\theta} \quad (5.27)$$

$$K_{\text{azimuthal}} = \frac{2\pi r^3}{Ll\sin\theta} \quad (5.28)$$

$$K_{\text{equatorial}} = \frac{2\pi r^3}{Ll} \quad (5.29)$$

The dipole field dies down inversely as the cube of distance. Therefore, the measured potential (ΔV) reduces downwards rapidly with dipole length r . The figure in Naskar and Das (2009) shows the variation of potential with varying dipole length and dipole separation. Koefoed (1979) had shown the connecting relation between the apparent resistivity obtained for the Schlumberger configuration and that obtained using the parallel dipole configuration:

$$\rho_{\text{app.dip.par}} = \rho_{\text{app.Sch}} - \frac{\cos^2\theta}{3\cos^2\theta - 1} r \frac{\partial \rho_{\text{app.Sch}}}{\partial r}$$

where $\rho_{\text{app.dip.par}}$ is the apparent resistivity obtained using a parallel dipole configuration, $\rho_{\text{app.Sch}}$ is the apparent resistivity obtained using the Schlumberger electrode configuration, θ is the dipole angle, and r is the half-electrode separation, that is, $\frac{AB}{2} = r$.

Since a parallel dipole becomes an equatorial dipole at $\theta = \pi/2$, one gets the theoretical justification of attaching the equatorial dipole data with the Schlumberger data for deeper probing. Normally, the shallower part of the probing is done with Schlumberger and the deeper part with the equatorial dipole array. The interconvertibility of the dipole apparent resistivity to the Schlumberger apparent resistivity is discussed in Bhattacharya and Patra (1968) and Koefoed (1979).

The dipole method of measuring resistivity is practical for use in studying the electrical properties of the Earth's crust. In such studies, an electrical current is driven into the ground with a pair of electrodes spaced a few kilometres apart, and the electric field caused by this current is mapped to distances as great as 100 km. The principal advantage of the dipole method is the relatively small

horizontal profiling were made to delineate a fault zone where steam can be tapped for energy. In the Bakreswar geothermal area, Birbhum District, West Bengal, electrical soundings and horizontal profiling were made to delineate a fault zone (Naskar, 2017). Resistivity sounding is not always a very efficient tool to map the vertical/subvertical fault planes, and hence, collinear dipole–dipole traversing is effective for the purpose (Naskar et al., 2002).

5.7 EFFECT OF INHOMOGENEOUS GROUND

So far, we have considered current flow and potential in and over homogeneous ground, a situation which is extremely rare in the field and which would be of no practical significance anyway. What we want to detect is the presence of anomalous conductivity in various forms, such as lumped (three-dimensional) bodies, dykes, faults, and vertical or horizontal contacts between beds. The resistivity method is most suitable for outlining horizontal beds and vertical contacts, less useful on bodies of irregular shape.

5.7.1 DISTORTION OF CURRENT FLOW AT A PLANE INTERFACE

Consider two homogeneous media of resistivities ρ_1 and ρ_2 separated by a plane boundary as in Figure 5.22 (Kearey and Brooks, 2002). Suppose that a current of density J_1 is flowing in medium 1 in such a direction as to meet the boundary at an angle θ_1 to the normal. To determine the direction of this current in medium 2, we recall the conditions given in eq. 5.11; using Ohm's law to express these results in terms of the current density, we obtain

$$J_{x_1}\rho_1 = J_{x_2}\rho_2 \text{ and } J_{z_1} = J_{z_2}$$

Dividing these expressions, we have

$$\rho_1 \left(\frac{J_{x_1}}{J_{z_1}} \right) = \rho_2 \left(\frac{J_{x_2}}{J_{z_2}} \right) \Rightarrow \rho_1 \tan\theta_1 = \rho_2 \tan\theta_2 \Rightarrow \frac{\tan\theta_2}{\tan\theta_1} = \frac{\rho_1}{\rho_2} \quad (5.31)$$

Thus, the current lines are bent in crossing the boundary. If $\rho_1 < \rho_2$, they will be bent toward the normal, and vice versa.

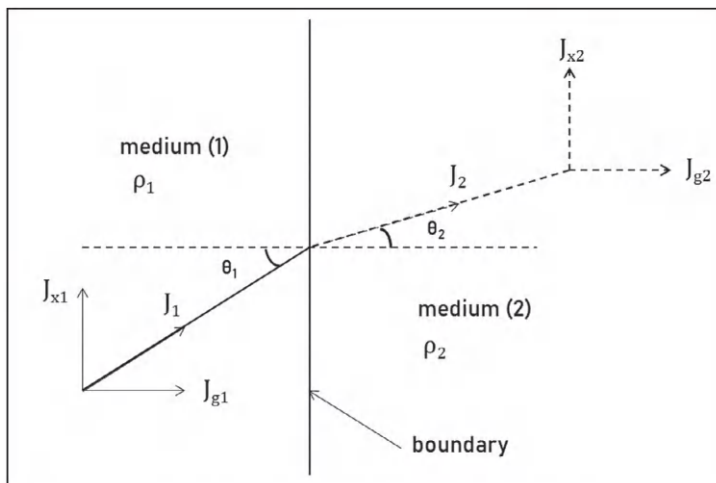


FIGURE 5.22 Distortion of current flow at a plane boundary when $\rho_1 < \rho_2$.

5.7.2 DISTORTION OF POTENTIAL AT A PLANE INTERFACE

Clearly, if the current flow is distorted in passing from a medium of one resistivity into another, the equipotentials also will be distorted. It is possible to determine the potential field mathematically by solving Laplace's equation for the appropriate boundary conditions or by integrating it directly. Both methods require considerable mathematics. A much simpler approach employs electrical images, in analogy with geometrical optics. The use of images is valid in solving only a limited number of potential problems, including the plane boundary and the sphere (Bewley, 1963).

The analogy between the electrical situation and optics is based on the fact that current density, like light ray intensity, decreases with the inverse square of distance from a point source in a medium of resistivity ρ_1 separated from an adjacent medium ρ_2 by a plane boundary.

In optics, the analogous case would be a point source of light in one medium separated from another by a semitransparent mirror, having reflection and transmission coefficients k and $(1 - k)$. Then, the light intensity at a point in the first medium is partly due to the point source and partly to its image in the second medium, the latter effect diminished by reflection from the mirror. On the other hand, the intensity at a point in the second medium is due only to the source in the first, diminished by transmission through the mirror (Figure 5.23a). If we replace the point source of light by a point source of current, and the light intensity at a point by potential, the problem is now in the electrical domain. From Figure 5.23b we see that the potential at P in the first medium is

$$V = \frac{I\rho_1}{4\pi} \left(\frac{1}{r_1} + \frac{k}{r_2} \right) \quad (5.32)$$

And in the second medium at P it is

$$V' = \frac{I\rho_2}{4\pi} \left(\frac{1-k}{r_3} \right) \quad (5.33)$$

Applying the boundary condition of eq. 5.11, these potentials must be equal at the interface when $r_1 = r_2 = r_3$. Thus, we have

$$\frac{\rho_1}{\rho_2} = \frac{1-k}{1+k} \Rightarrow k = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1} \quad (5.34)$$

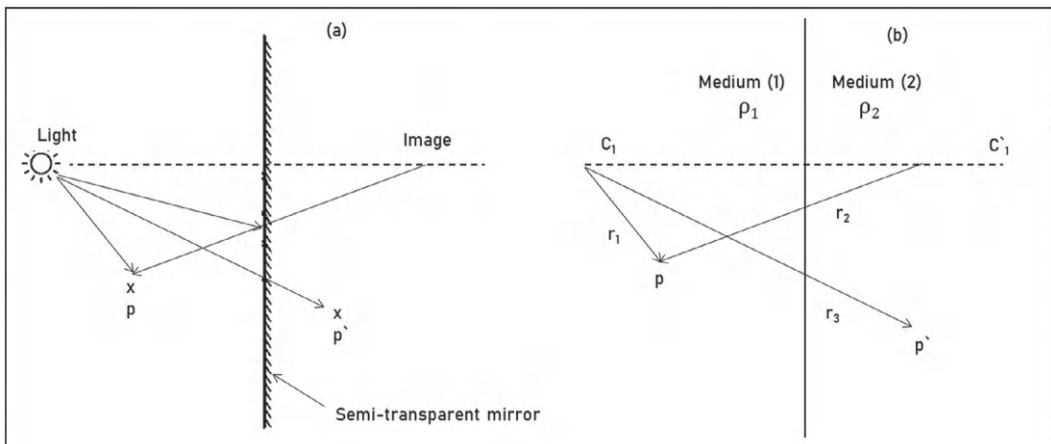


FIGURE 5.23 Analogy between optical and electrical images: (a) optical image, and (b) electrical image.

Source: Kearey and Brooks (2002).

In this expression, k is a reflection coefficient whose value lies between ± 1 , depending on the relative resistivities in the two media.

If the second layer is a pure insulator ($\rho_2 = \infty$), $k = +1$; and if the second layer is a perfect conductor ($\rho_2 = 0$), then $k = -1$. When $\rho_2 = \rho_1$, no electrical boundary exists and, as might be expected, $k = 0$.

In the preceding treatment, eqs. 5.32 and 5.33 are valid for a single point source of current ($+I$). The effect of a ($-I$) current point source (second-current electrode) can be calculated in a similar manner, and the total potential in each medium can be determined by superposition.

Now, imagine that the conditions shown in Figure 5.23 are modified. If the boundary separating the two layers is horizontal (instead of vertical), the problem is complicated by the fact that there are now three layers, the air forming the third layer (Figure 5.24). In these situations, an infinite number of 'images' is produced of a source placed at the ground surface. The original image C_1' at depth $2h$ below the surface is reflected in the surface boundary to give an image C_1'' at a distance $2h$ above C_1 . This second image, reflected in the lower boundary, produces a third C_1''' at a depth $4h$, and so on. The situation is somewhat analogous to that of a light source (e.g. a candle) placed between two parallel mirrors.

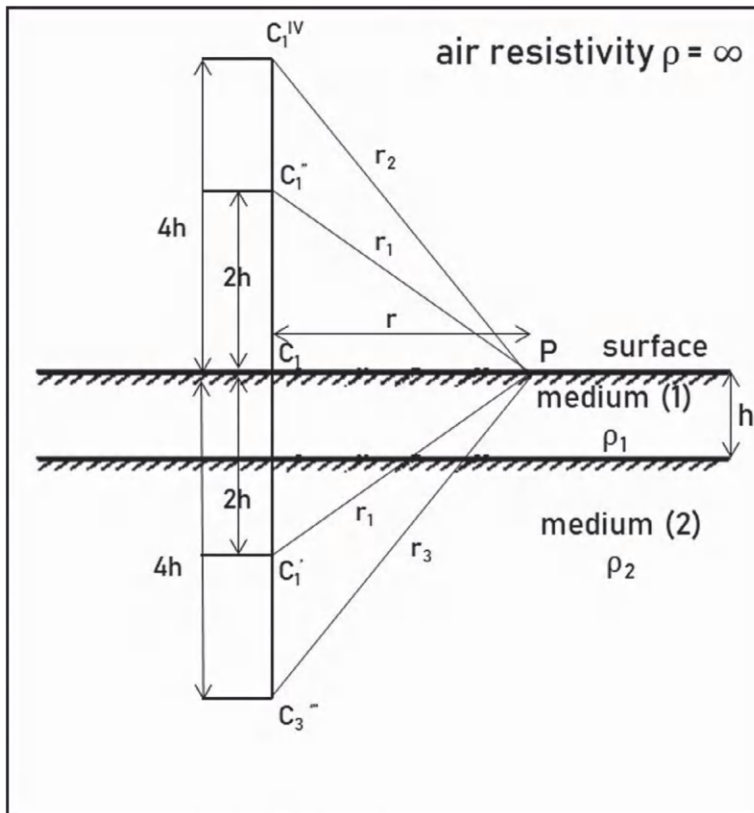


FIGURE 5.24 Quantities involved in the calculation of the potential due to a single-current point source above a horizontal discontinuity separating two media of resistivity ρ_1 and ρ_2 . C_1' , C_1'' , C_1''' , and C_1^{IV} are the nearer images of the current source C_1 , successively produced by reflections at the $\frac{\rho_1}{\rho_2}$ interface and the $\frac{\rho_2}{\rho_1}$ air interface.

The effect of each successive image on the potential at P is reduced by the reflection coefficient between the boundaries. For the current source and its first image belowground, the potential as eq. 5.32 is:

$$V' = \frac{I\rho_1}{2\pi} \left(\frac{1}{r} + \frac{k}{r_1} \right)$$

The effect of the second image at C_1'' , 2 h aboveground, is

$$V'' = \frac{I\rho_1}{2\pi} \left(\frac{k \times k_a}{r_1} \right)$$

where k_a is the reflection coefficient at the surface boundary. Because ρ_a is essentially infinite, this coefficient is unity. Therefore:

$$V' + V'' = \frac{I\rho_1}{2\pi} \left(\frac{1}{r} + \frac{2k}{r_1} \right)$$

The potential due to the third image C_1''' , 4 h belowground, will be further reduced, as will that of its image 4 h above ground. Hence:

$$V''' + V^{IV} = \frac{I\rho_1}{2\pi} \left(\frac{k \times k}{r_2} + \frac{k \times k \times k_a}{r_2} \right) = \frac{I\rho_1}{2\pi} \left(\frac{2k^2}{r_2} \right)$$

The resulting potential at, for instance, a point P on the ground surface is the sum of the potentials due to the current source C_1 (of strength I) and its infinite series of images. Using the notation of Figure 5.24 (Telford et al., 1990) and employing eq. 5.32 give

$$V_P = \frac{I\rho_1}{2\pi} \left(\frac{1}{r} + \frac{2k}{r_1} + \frac{2k^2}{r_2} + \dots + \frac{2k^n}{r_n} + \dots \right) = \frac{I\rho_1}{2\pi} \left(\frac{1}{r} + 2 \sum_{n=1}^{\infty} \frac{k^n}{r_n} \right) \quad (5.35)$$

where $r_n = \left\{ r^2 + (2nh)^2 \right\}^{\frac{1}{2}}$ and k is the dimming factor in eq. 5.33. The parameter h is defined in the figure.

The series is convergent because k (absolute value) < 1 , and the denominator increases indefinitely. The first term in the parentheses of eq. 5.34 refers to the normal potential arising if the subsurface were homogeneous; the remaining terms, expressed as an infinite series, define the disturbing potential.

Eq. (5.35) relates the potential of a single electrode to the resistivity of the upper layer in terms of the electrode spacing, the depth to the interface, and the resistivity contrast between the two beds. We want this expression in the form of an apparent resistivity, which would be measured by a four-electrode system. Using Figure 5.10 and eqs. 5.15, 5.34, and 5.35, it is possible to compute the potential difference (ΔV) between two points, resulting from two current electrodes, by the addition and subtraction of their contributions to the potential at these points (taking $AM = r_1$, $BM = r_2$, $AN = r_3$, and $BN = r_4$).

$$\Delta V = V_1 - V_2$$

$$= \frac{I\rho_1}{2\pi} \left[\left(\frac{1}{r_1} - \frac{1}{r_2} - \frac{1}{r_3} + \frac{1}{r_4} \right) + 2 \sum_{i=1}^{\infty} k^n \left\{ \frac{1}{\left(r_1^2 + 4n^2 h^2 \right)^{\frac{1}{2}}} - \frac{1}{\left(r_2^2 + 4n^2 h^2 \right)^{\frac{1}{2}}} \right. \right. \\ \left. \left. - \frac{1}{\left(r_3^2 + 4n^2 h^2 \right)^{\frac{1}{2}}} + \frac{1}{\left(r_4^2 + 4n^2 h^2 \right)^{\frac{1}{2}}} \right\} \right]$$

For the Wenner electrode arrangement (Figure 5.14) with spacing a and $r_1 = r_4 = a$, $r_2 = r_3 = 2a$, the computed expression for ΔV between the two points P_1 and P_2 becomes

$$\Delta V = \frac{I\rho_1}{2\pi} \left[1 + \sum_{n=1}^{\infty} \frac{4k^n}{\left\{ 1 + \left(\frac{2nh}{a} \right)^2 \right\}^{\frac{1}{2}}} - \sum_{n=1}^{\infty} \frac{4k^n}{\left\{ 4 + \left(\frac{2nh}{a} \right)^2 \right\}^{\frac{1}{2}}} \right]$$

$$\Delta V = \frac{I\rho_1}{2\pi a} (1 + 4S_w) \quad (5.36)$$

$$\rho_a = \rho_1 (1 + 4S_w)$$

$$\text{where } S_w = \sum_{n=1}^{\infty} k^n \left[\frac{1}{\left\{ 1 + \left(\frac{2nh}{a} \right)^2 \right\}^{\frac{1}{2}}} - \frac{1}{\left\{ 4 + \left(\frac{2nh}{a} \right)^2 \right\}^{\frac{1}{2}}} \right]$$

Exploiting the linear relationship between the resistivity transform and the apparent resistivity functions Ghosh (1971) shows that only a simple filtering operation is required to transform the $T(\lambda)$ values for a layered sequence into ρ_a values. By using a small number of fixed coefficients (calculated once for all), it is possible to make very rapid computations of theoretical ρ_a curves for a multilayered resistivity distribution. The problem is recast in the form

$$\rho_a = \sum T(\lambda) f_k$$

where the coefficients of a moving average filter are represented by f_k . The linear filtering approach can be effectively used to transform resistivity sounding data obtained with one type of electrode configuration into another (Kumar and Das, 1977; Koefoed, 1979).

A good review of the theory of linear filters pertaining to the calculation of resistivity curves can be found in Koefoed's (1979) monograph. This work also includes programs usable on microcomputers and pocket calculators for the computation of the Schlumberger theoretical apparent resistivity ρ_a curves for a layered structure (1-D forward modelling). Computation of apparent resistivities due to 2-D and 3-D structures is more complicated than that for the 1-D model.

5.8 RESISTIVITY SURVEYING

To get an idea of what is measured with resistivity methods, we have to first look at how the electricity flows through the rocks. The current is injected into the ground by a pair of electrodes—metal sticks pushed into the ground. The positive- and negative-current electrodes are often denoted as C_1 and C_2 , or A and B . The current flows between the electrodes using the easiest path, which is the path with the lowest resistance. As the resistance decreases with the increasing diameter of the wire, the current paths spread downwards and sideways (Figure 5.11).

Nevertheless, the highest concentration of the current is near the surface. In uniform ground, only about 30% of the current penetrates deeper than is the separation of the current electrodes. This also implies that the depth of penetration of the current and a volume of rock sampled depends on the distance between the current electrodes (Musset and Khan, 2000). The further the electrodes are placed, the deeper the penetration; however, a larger volume of rock is sampled, and hence, the survey is less detailed.

The potential difference is measured by another pair of electrodes with a voltmeter connected to them (Musset and Khan, 2000). These potential electrodes are usually called P_1 and P_2 , or M and N . The potential difference measured between the potential electrodes is usually denoted ΔV .

The voltage applied to the current electrodes generally depends on the separation of electrodes and resistivities of the subsurface but usually is between 100 and 300 V for the near-surface surveys. The current is, however, usually less than 1 A, and the potential difference read is in millivolts. The potential difference readings are heavily affected by ions concentrating on the electrodes, creating an additional potential (and often also time variant). To diminish this effect, a less-polarizing metal could be used for the electrodes (e.g. a stainless steel or brass), or the current could be reversed a few times a second to disperse the ions. If very precise potential difference readings are needed (the case of measuring spontaneous or induced polarization), the so-called nonpolarizing electrodes are used. They consist of a metal immersed in a saturated solution of its own salt, such as Cu in a CuSO_4 , contained in a porous, permeable ceramic pot. The solution slowly leaks through the pores and ensures a proper grounding.

Up to now, we have been examining the current paths in the homogeneous media only. If there would be interfaces of layers with different resistivities, the flow paths would bend or refract similarly to light or seismic ray paths (Musset and Khan, 2000; in Telford et al., 1990; in Musset and Khan, 2000—three figures). They refract towards the normal when crossing into a rock with higher resistivities, and conversely in a rock with lower resistivities—not the same as Snell's law, and opposite behaviour from seismic refraction. The relationship is $\rho_1 \tan \theta_1 = \rho_2 \tan \theta_2$. The refraction changes the distribution of the current when compared with the uniform ground and distortion of potential at a plane interface occurs and hence changes also the $\frac{\Delta V}{I}$ ratio; it is possible to measure changes of resistivity.

The resistivity measured over an inhomogeneous media is an overall resistivity combined from the resistivities of all layers and bodies affecting the flow paths. This resistivity is called the apparent resistivity, ρ_a , and is the resistivity that uniform ground giving the same ($\frac{\Delta V}{I}$) with the same electrode separations would have.

There are two basic modes of resistivity surveying: sounding and profiling. *Sounding* benefits from the fact that the depth of the penetration increases with a distance of the current electrodes. Hence, repeated measurements on one place with increasing distance of current electrodes measures different depth levels, and a vertical profile of a subsurface could be derived (similar to a borehole). The depth of penetration depends on the resistivity values of rocks encountered; however, as a first rough estimate, one-fourth of the distance between the current electrodes could be considered as a depth estimate. The technique is best adapted to determining depth and resistivity for flat-lying layered rock structures, such as sedimentary beds, or the depth to the water table. The Schlumberger configuration is most commonly used for vertical electrical sounding (VES) investigations.

In electrical sounding with the Wenner, Schlumberger, or dipole–dipole arrays, the respective electrode spacing a , $\frac{AB}{2}$, or r is increased at successive logarithmic intervals, and the value of the

appropriate apparent resistivity, ρ_W , ρ_S , and ρ_D , is plotted as a function of the electrode spacing on a logarithmic coordinate paper. The curve of $\rho_a = f(a, \frac{AB}{2}, \text{ or } r)$ is called an electrical sounding curve.

The profiling uses the same interelectrode distances for all measurements, but the whole array moves along the profile. Thus, lateral changes of resistivities could be mapped. The value of apparent resistivity is plotted, generally, at the geometric centre O of the electrode array. Maximum apparent resistivity anomalies are obtained by orienting the profiles at right angles to the strike of the geologic structure. The results are presented as apparent resistivity profiles (Zohdy et al., 1964). In making horizontal profiles, it is recommended that at least two different electrode spacings be used, in order to aid in distinguishing the effects of shallow geologic structures from the effects of deeper ones (Zohdy, 1964). In the figure in Zohdy (1964), the effect of shallow geologic features is suppressed on the profile made with the larger spacing, whereas the effect of deeper features is retained. The interpretation of horizontal profiling data is generally qualitative, and the primary value of the data is to locate geologic structures such as buried stream channels, veins, and dykes. Quantitative interpretation can be obtained by making a sufficient number of profiles with different electrode spacings and along sets of traverses of different azimuths. Best interpretative results are obtained generally from a combination of horizontal profiling and electrical sounding data. Finally, a combination of both modes—several measurements on a profile with different interelectrode distance—maps both the lateral and vertical changes in resistivities producing the 2-D resistivity cross section.

5.9 COMPARISON OF WENNER, SCHLUMBERGER, AND DIPOLE–DIPOLE MEASUREMENTS

The Schlumberger and the Wenner electrode arrays are the two most widely used arrays in resistivity prospecting. These are the essential differences between three arrays:

- In the Schlumberger array, the distance between the potential electrodes MN is small and is always kept equal to, or smaller than, one-fifth the distance between the current electrodes AB , that is, $AB \geq 5 MN$. In the Wenner array, AB is always equal to $3 MN$.
- In a Schlumberger sounding, the potential electrodes are moved only occasionally, whereas in a Wenner sounding, they and the current electrodes are moved after each measurement.

With the invention of dipole–dipole arrays and their use in the Soviet Union and the United States, their following advantages over the Schlumberger array became recognized:

- Relatively short AB and MN lines are used to explore large depths, which reduce field labour and increase productivity.
- Problems of current leakage (Dakhnov, 1963; Zohdy, 1968b) are reduced to a minimum.
- Bilateral investigations are possible, and therefore, more detailed information on the direction of dip of electrical horizons is obtainable.
- Problems of inductive coupling and associated errors are minimized.

Among the disadvantages of dipole methods are:

- The requirement of a large generator to provide ample amounts of current, especially in deep exploration.
- Special knowledge and special theoretical developments and materials are required to interpret most of the data obtained by dipole–dipole arrays. Generally, one cannot use the experience gained in using Schlumberger or Wenner arrays to obtain or to interpret dipole sounding data in a straightforward way.

5.10 PROBLEM OF DEFINING PROBING DEPTH

A favourite rule of thumb in electrical prospecting is that the electrode spacing is equal to the depth of probing. This rule of thumb is wrong and leads to erroneous interpretations. Its origin probably stems from the fact that when using direct current in probing a homogeneous and isotropic semi-infinite medium, there is a definite relation between the spacing AB separating the current electrodes and the depth to which any particular percentage of the current penetrates. For example, 50% of the current penetrates to a depth equal to $\frac{AB}{2}$, and 70% to a depth equal to AB . Therefore, the greater the current electrode separation, the greater the amount of current that penetrates a given depth. This relation is governed by eq. 5.37 (Weaver, 1929; Jakosky, 1950):

$$\frac{I_s}{I_t} = \frac{2}{\pi} \tan^{-1} \left(\frac{2z}{AB} \right) \quad (5.37)$$

where I_s is the current confined between depths 0 and z , I_t is the total current penetrating the ground, and AB is the distance separating the current electrodes.

This current–depth relation for a homogeneous and isotropic Earth cannot be used as a general rule of thumb to establish a so-called ‘depth of penetration’ or ‘probing depth’ that also applies to a stratified or an inhomogeneous Earth. For an inhomogeneous medium, the percentage of the total current that penetrates a given depth z depends not only upon the electrode separation but also upon the resistivities of the Earth layers. This fact was discussed by Muskat (1933), Muskat and Evinger (1941), Evjen (1938), Orellana (1960, 1961), and others.

5.11 ADVANTAGES OF USING LOGARITHMIC COORDINATES

Electrical sounding data should be plotted on logarithmic coordinates with the electrode spacing on the abscissa and the apparent resistivity on the ordinate. The advantages of plotting the sounding data on logarithmic coordinates are:

- Field data can be compared with precalculated theoretical curves for given Earth models (curve matching procedure).
- The form of an electrical sounding curve does not depend on the resistivity and thickness of the first layer, provided that the ratios $\frac{\rho_2}{\rho_1}, \frac{\rho_3}{\rho_1}, \dots, \frac{\rho_n}{\rho_1}$, and the $\frac{h_2}{h_1}, \frac{h_3}{h_1}, \dots, \frac{h_n}{h_1}$, remain constant from model to model, where $\rho_1, \rho_2, \rho_3, \dots, \rho_n$ are the resistivities and $h_1, h_2, h_3, \dots, h_n$ are the thicknesses of the first, second, third, and n th layers, respectively.
- The use of logarithmic coordinates, on the one hand, suppresses the effect of variations in the thickness of layers at large depths, and it also suppresses variations of high resistivity values. On the other hand, it enhances the effect of variations in the thickness of layers at shallow depths, and it enhances the variations of low resistivity values.
- The wide spectrum of resistivity values measured under different field conditions and the large electrode spacings, necessary for exploring the ground to moderate depths, make the use of logarithmic coordinates a logical choice.

5.12 GEOELECTRIC PARAMETERS

A *geologic* section differs from a *geoelectric* section when the boundaries between geologic layers do not coincide with the boundaries between layers characterized by different resistivities. Thus, the electric boundaries separating layers of different resistivities may or may not coincide with boundaries separating layers of different geologic age or different lithologic composition.

A *geolectric layer* is described by two fundamental parameters: its resistivity ρ_i and its thickness h_i , where the subscript i indicates the position of the layer in the section ($i = 1$ for the uppermost layer). Other geoelectric parameters are derived from its resistivity and thickness. These are:

- Longitudinal unit conductance: $S_i = \frac{h_i}{\rho_i}$ (5.38)

- Transverse unit resistance: $T_i = h_i \rho_i$ (5.39)

- Longitudinal resistivity: $\rho_L = \frac{h_i}{S_i}$ (5.40)

- Transverse resistivity: $\rho_t = \frac{T_i}{h_i}$ (5.41)

- Anisotropy: $\lambda = \sqrt{\frac{\rho_t}{\rho_L}}$ (5.42)

For an isotropic layer, $\rho_t = \rho_L$ and $\lambda = 1$. These secondary geoelectric parameters are particularly important when they are used to describe a geoelectric section consisting of several layers. For n layers, the total longitudinal unit conductance is $S = \sum_{i=1}^n \frac{h_i}{\rho_i} = \frac{h_1}{\rho_1} + \frac{h_2}{\rho_2} + \dots + \frac{h_n}{\rho_n}$, the total

transverse unit resistance is $T = \sum_{i=1}^n h_i \rho_i = h_1 \rho_1 + h_2 \rho_2 + \dots + h_n \rho_n$, the average longitudinal

resistivity is $\rho_L = \frac{H}{S} = \frac{\sum_{i=1}^n h_i}{\sum_{i=1}^n \frac{h_i}{\rho_i}}$, the average transverse resistivity is $\rho_t = \frac{T}{H} = \frac{\sum_{i=1}^n h_i \rho_i}{\sum_{i=1}^n h_i}$, and the

anisotropy is $\lambda = \sqrt{\frac{\rho_t}{\rho_L}} = \frac{\sqrt{TS}}{H}$. The parameters S , T , ρ_L , ρ_t , and λ are derived from the consideration

of a column of unit square cross-sectional area ($1 \times 1 \text{ m}^2$) cut out of a group of layers of infinite lateral extent (Figure 5.25) (Zohdy, 1964). If current flows vertically only through the column, then the layers in the column will behave as resistors connected in series, and the total resistance of the column of unit cross-sectional area will be

$$R = R_1 + R_2 + R_3 + \dots + R_n$$

$$\text{or } R = \rho_1 \frac{h_1}{1 \times 1} + \rho_2 \frac{h_2}{1 \times 1} + \dots + \rho_n \frac{h_n}{1 \times 1} = \sum_{i=1}^n \rho_i h_i = T$$

The symbol T is used instead of R to indicate that the resistance is measured in a direction transverse to the bedding, and also because the dimensions of this 'unit resistance' are usually expressed in ohm-m² instead of ohms.

If the current flows parallel to the bedding, the layers in the column will behave as resistors connected in parallel and the conductance will be

$$S = \frac{1}{R} = \frac{1}{R_1} + \frac{1}{R_2} + \frac{1}{R_3} + \dots + \frac{1}{R_n}$$

$$\text{or } S = \frac{1 \times h_1}{\rho_1 \times 1} + \frac{1 \times h_2}{\rho_2 \times 1} + \frac{1 \times h_3}{\rho_3 \times 1} + \dots + \frac{1 \times h_n}{\rho_n \times 1} \Rightarrow S = \frac{h_1}{\rho_1} + \frac{h_2}{\rho_2} + \dots + \frac{h_n}{\rho_n}$$

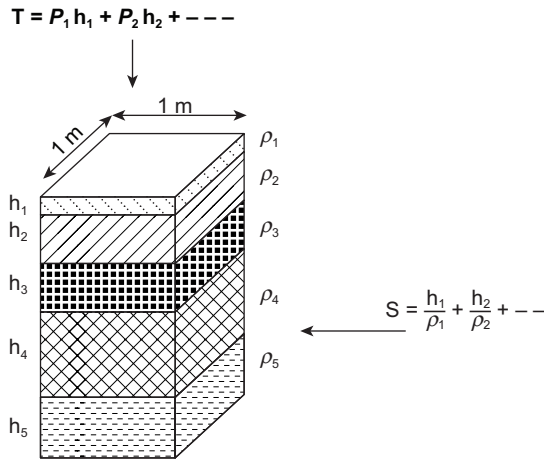


FIGURE 5.25 Columnar prism used in defining geoelectric parameters of a section. Patterns are arbitrary (ρ = resistivity, h = thickness, S = total longitudinal conductance, T = total transverse resistance).

The dimensions of the longitudinal unit conductance are $\frac{\text{m}}{\text{ohm} - \text{m}} = \frac{1}{\text{ohm}} = \text{mho}$. It is interesting to note that the quantity $S_i = \frac{h_i}{\rho_i} = \sigma_i h_i$, where σ_i is the conductivity (inverse of resistivity), is analogous to transmissivity $T_i = K_i b_i$ used in groundwater hydrology, where K_i is the hydraulic conductivity of the i^{th} layer and b_i is its thickness. The parameters T and S were named the ‘Dar Zarrouk’ parameters by Mailliet (1947). In the interpretation of multilayer electrical sounding curves, the evaluation of S or T is sometimes all that can be determined uniquely. The study of the parameters S , T , ρ_L , ρ_T , and λ is an integral part of the analysis of electrical sounding data and also is the basis of important graphical procedures (for example, the auxiliary point method) for the interpretation of electrical sounding curves (Kalenov, 1957; Orellana and Mooney, 1966; Zohdy, 1965a).

5.13 VERTICAL ELECTRICAL SOUNDING

5.13.1 TYPES OF ELECTRICAL SOUNDING CURVES OVER HORIZONTALLY STRATIFIED MEDIA

A two-layer situation is encountered often in electrical prospecting, for example, when a conducting overburden overlies a resistive basement. It is also common in environmental applications, when the conducting water table lies under drier, more resistive soil or rocks. Before the advent of portable computers, two-layer cases were interpreted with the aid of *characteristic curves*. These theoretical curves, calculated for a particular four-electrode array, take into account the change in depth penetration when current lines cross the boundary to a layer with different resistivity. The electrical boundary conditions require continuity of the component of current density J normal to the interface and of the component of electric field E tangential to the interface. At a boundary, the current lines behave like optical or seismic rays and are guided by similar laws of reflection and refraction. For example, if θ is the angle between a current line and the normal to the interface, the electrical

‘law of refraction’ is $\frac{\tan \theta_1}{\tan \theta_2} = \frac{\rho_2}{\rho_1}$. In a set of characteristic curves, the apparent resistivity ρ_a is

normalized by the resistivity ρ_1 of the upper layer, and the electrode spacing is expressed as a multiple of the layer thickness. The shape of the curve of apparent resistivity versus electrode spacing depends on the resistivity contrast between the two layers, and a family of characteristic curves is

calculated for different ratios of $\left(\frac{\rho_2}{\rho_1}\right)$ (Zohdy et al., 1964; Orellana and Mooney, 1966). The resistivity contrast is conveniently expressed by a k factor, defined as $k = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}$. The k factor ranges between -1 and $+1$, as the resistivity ratio $\left(\frac{\rho_2}{\rho_1}\right)$ varies between 0 and ∞ . The characteristic curves, drawn as full logarithmic plots on a transparent overlay, are compared graphically with the field data to find the best-fitting characteristic curve. The comparison yields the resistivities ρ_1 and ρ_2 of the upper and lower layers, respectively, and the layer thickness (d). Although characteristic curves can also be computed for the interpretation of structures with multiple horizontal layers, modern VES analyses take advantage of the flexibility offered by small computers with graphic outputs on which the apparent resistivity curves can be assessed visually. The first step in the analysis consists of classifying the shape of the vertical sounding profile. If the ground is composed of a single homogeneous and isotropic layer of infinite thickness and finite resistivity, then, irrespective of the electrode array used, the apparent resistivity curve will be a straight horizontal line whose ordinate is equal to the true resistivity ρ_1 of the semi-infinite medium.

If the ground is composed of two layers, a homogeneous and isotropic first layer of thickness h_1 and resistivity ρ_1 , underlain by an infinitely thick substratum ($h_2 = \infty$) of resistivity ρ_2 , then the sounding curve begins, at small electrode spacings, with a horizontal segment ($\rho_a \cong \rho_1$). As the electrode spacing is increased, the curve rises or falls, depending on whether $\rho_2 > \rho_1$ or $\rho_2 < \rho_1$ and on the electrode configuration used. At electrode spacings much larger than the thickness of the first layer, the sounding curve asymptotically approaches a horizontal line whose ordinate is equal to ρ_2 . The electrode spacing at which the apparent resistivity ρ_a asymptotically approaches the value ρ_2 depends on three factors: the thickness of the first layer (h_1), the value of the ratio $\frac{\rho_2}{\rho_1}$, and the type of electrode array used in making the sounding measurements.

For most electrode arrays, including the conventional Schlumberger, Wenner, dipole equatorial, and dipole polar arrays, when $\rho_1 > \rho_2 > \rho_3$, larger electrode spacings are required for ρ_a to be approximately equal to ρ_2 than when $\frac{\rho_2}{\rho_1} < 1$. The figure in Zohdy et al. (1964) shows a comparison between two Schlumberger sounding curves obtained over two-layer Earth models in which $h_1 = 1$ m, $\frac{\rho_2}{\rho_1} = 10$, and $\frac{\rho_2}{\rho_1} = 0.1$.

The apparent resistivity curve for a three-layer structure (of resistivities ρ_1 , ρ_2 , and ρ_3 and thicknesses h_1 , h_2 , and $h_3 = \infty$) generally has one of four typical shapes, determined by the vertical sequence of resistivities in the layers (Figure 5.26) (Zohdy, 1965a).

The type K curve rises to a maximum then decreases, indicating that the intermediate layer has higher resistivity than the top and bottom layers. The type H curve shows the opposite effect; it falls to a minimum, then increases again due to an intermediate layer that is a better conductor than the top and bottom layers. The type A curve may show some changes in gradient, but the apparent resistivity generally increases continuously with increasing electrode separation, indicating that the true resistivities increase with depth from layer to layer. The type Q curve exhibits the opposite effect; it decreases continuously along with a progressive decrease of resistivity with depth. These are:

$\rho_1 < \rho_2 > \rho_3$: K-type section

$\rho_1 > \rho_2 < \rho_3$: H-type section

$\rho_1 < \rho_2 < \rho_3$: A-type section

$\rho_1 > \rho_2 > \rho_3$: Q-type section

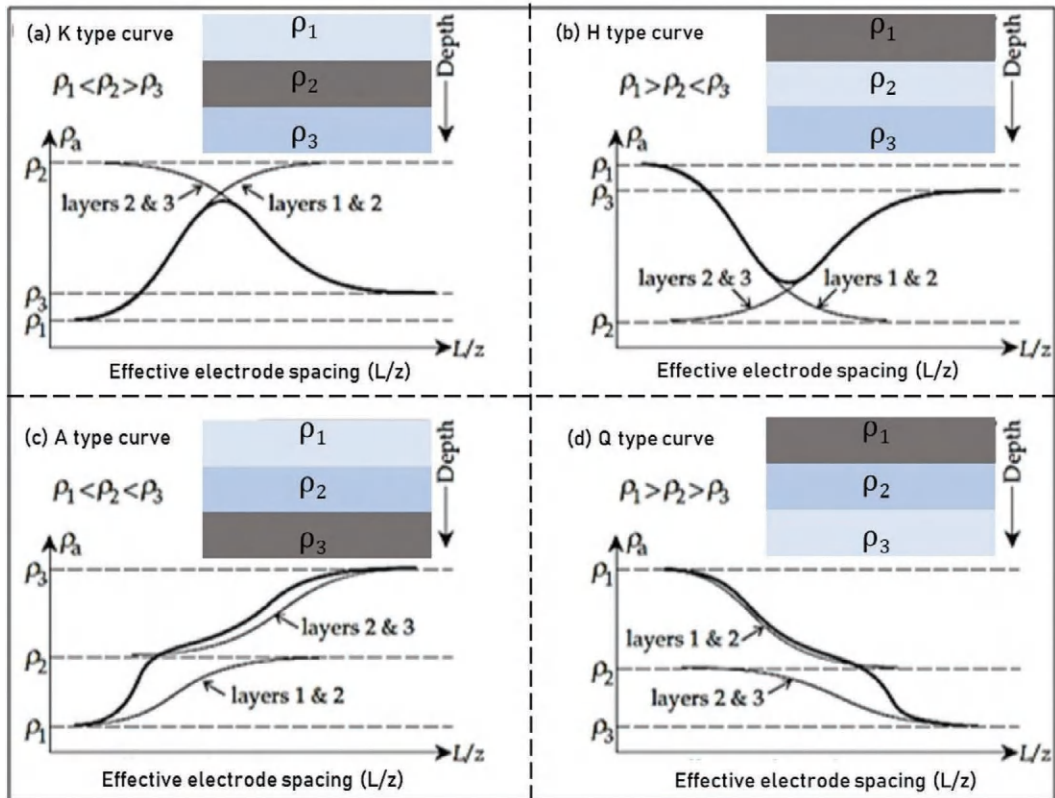


FIGURE 5.26 The four common shapes of apparent resistivity curves for a layered structure consisting of three horizontal layers.

Once the observed resistivity profile has been identified as of K, H, A, or Q type, the next step is equivalent to one-dimensional *inversion* of the field data. The technique involves iterative procedures that would be very time-consuming without a fast computer. The method assumes the equations for the theoretical response of a multilayered ground. Each layer is characterized by its thickness and resistivity, each of which must be determined. A first estimate of these parameters is made for each layer, and the predicted curve of apparent resistivity versus electrode spacing is computed. The discrepancies between the observed and theoretical curves are then determined point by point. The layer parameters used in the governing equations are next adjusted, and the calculation is repeated with the corrected values, giving a new predicted curve to compare with the field data. Using modern computers, the procedure can be reiterated rapidly until the discrepancies are smaller than a predetermined value.

If the ground is composed of more than three horizontal layers of resistivities $\rho_1, \rho_2, \rho_3, \dots, \rho_n$ and thicknesses $h_1, h_2, h_3, \dots, h_n = \infty$, the geoelectric section is described in terms of the relationship between the resistivities of the layers, and the letters *H*, *A*, *K*, and *Q* are used, in combination, to indicate the variation of resistivity with depth. In four-layer geoelectric sections, there are eight possible relations between ρ_1, ρ_2, ρ_3 , and ρ_4 :

$$\rho_1 > \rho_2 < \rho_3 < \rho_4 \dots \dots \dots, \text{HA-type section}$$

$$\rho_1 > \rho_2 < \rho_3 > \rho_4 \dots \dots \dots, \text{HK-type section}$$

$\rho_1 < \rho_2 < \rho_3 < \rho_4 \dots\dots\dots$, AA-type section

$\rho_1 < \rho_2 < \rho_3 > \rho_4 \dots\dots\dots$, AK-type section

$\rho_1 < \rho_2 > \rho_3 < \rho_4 \dots\dots\dots$, KH-type section

$\rho_1 < \rho_2 > \rho_3 > \rho_4 \dots\dots\dots$, KQ-type section

$\rho_1 > \rho_2 > \rho_3 < \rho_4 \dots\dots\dots$, QH-type section

$\rho_1 > \rho_2 > \rho_3 > \rho_4 \dots\dots\dots$, QQ-type section

Examples of Schlumberger electrical sounding curves for three of these eight types of four-layer models are shown in Zohdy et al. (1964). The inversion method is equivalent to matching automatically the observed and theoretical curves. A one-dimensional analysis accommodates only the variations of resistivity and layer thickness with depth. The response of a vertically layered structure has an analytical solution, so efficient inversion algorithms can be established. In recent years, procedures have been proposed that also take into account lateral heterogeneities. The response of two- or three-dimensional structures must be approximated by a numerical solution, based on the finite difference or finite-element techniques. The number of unknown quantities increases, as do the computational difficulties of the inversion.

5.13.2 ELECTRICAL SOUNDING OVER Laterally INHOMOGENEOUS MEDIA

Lateral inhomogeneities in the ground affect resistivity measurements in different ways. The effect depends on:

- The size of the inhomogeneity with respect to its depth of burial
- The size of the inhomogeneity with respect to the size of the electrode array
- The resistivity contrast between the inhomogeneity and the surrounding media
- The type of electrode array used
- The geometric form of the inhomogeneity
- The orientation of the electrode array with respect to the strike of the inhomogeneity

The simplest type of a lateral inhomogeneity, from the geometric and mathematical points of view, is that of a vertical plane boundary separating two homogeneous and isotropic media of resistivities ρ_1 and ρ_2 . Although this Earth model is ideal and does not exist commonly in nature, its study serves to illustrate the general form of the resistivity anomaly to be expected over a large variety of more complicated lateral inhomogeneities.

The Wenner sounding curves for azimuth angles of 0° to 90° are shown in Figure 21, p. 30, in Zohdy et al. (1964; Zohdy, 1970). The Wenner curves are more complicated than the Schlumberger curves because a potential electrode crosses the contact. The effects of such things as dipping, vertical and horizontal contacts, and pipelines have been described in the literature for different electrode arrays (Kunetz, 1966; Alpin et al., 1966).

5.13.3 AMBIGUITIES OF SOUNDING CURVE INTERPRETATION

In VES, sounding curve modelling and interpretation are inextricably linked with the *principle of equivalence*. The principle expresses that a measured sounding curve is basically related with many physically equivalent models that may differ considerably.

A simple example is shown in Kirsch (2006). The sounding curve matches two four-layer models with alternating high and low resistivities. It is evident that, for the second, high-resistivity layer, a

doubling of the resistivity corresponds approximately with a halving of the thickness. For the third, low-resistivity layer, a halving of the resistivity can be compensated by halving of the thickness. According to this rule of equivalence, the maximum of a sounding curve can be modelled by the product of $T = m$ (thickness) $\times \rho$ (resistivity) only, and the minimum of a sounding curve by the ratio $S = \frac{m}{\rho}$ only. T (unit ohm-m²) and S (unit siemens or mho) are termed ‘transverse resistance’ and

‘longitudinal conductivity’, respectively. Consequently, the thickness of a layer cannot unambiguously be determined from VES without knowing its resistivity, and vice versa. Accordingly, the top of the 600 Ω -m layer (say, the top of an aquifer) in the example of Kirsch (2006) may exist at the depths of 31 m (model I) or 17 m (model II), depending on the resistivities of the overlying layers.

Equivalence is not restricted to this easily understandable situation. The sounding curve in Kirsch (2006) can be modelled by a five-layer as well as by a ten-layer Earth without, from the physical point of view, any preference. From Kirsch (2006), another important property of vertical electrical sounding curves is evident. By all appearances, only the minimum number of layers can be deduced, implying that thin layers are preferentially suppressed. As shown by the logarithmic scale of Kirsch (2006), this effect rapidly increases with increasing sounding depth. For the five-layer case, the comparable amplitudes of both sounding curve maxima are attributed to a thickness of 2 m (layer 2) at 1 m depth but to about 25 m (layer 4) at about 14 m depth. A rule of thumb indicates that a layer becomes clearly visible in a sounding curve if its thickness is comparable to its depth of deposition. The decreasing resolution with increasing depth is a basic drawback of the VES method since, at the same time, the problem of macro-anisotropy as a special case of equivalence increases.

Electrical anisotropy describes the effect that generally the resistivity of rocks is smaller parallel to bedding or schistosity compared with the resistivity transverse to these textures. Correspondingly, a longitudinal resistivity ρ_L and a transverse resistivity ρ_T are defined. For electrical soundings, current flow lines are curved so that transverse as well as longitudinal resistivities are encountered. From theory, modelling of an anisotropic bed (layer i) with thickness h_i in a sounding curve supplies a substitute resistivity ρ_i^* and substitute thickness h_i^* :

$$\rho_i^* = \sqrt{\rho_T \rho_L} \quad (5.43)$$

$$h_i^* = h_i \sqrt{\frac{\rho_T}{\rho_L}} \quad (5.44)$$

While for the substitute resistivity ($\rho_L < \rho_i^* < \rho_T$), the substitute thickness h_i^* is larger than the true thickness h_i by a multiplier of $\lambda = \sqrt{\frac{\rho_T}{\rho_L}}$, λ being the coefficient of anisotropy (Kirsch, 2006).

Geoelectric anisotropy is a matter of scale and of the resolution of a vertical resistivity section. While graphite, slate, or lake sediment may be termed *micro-anisotropic*, a sequence of well-defined electrically isotropic beds can behave macro-anisotropic if they are not resolved in a geoelectric sounding curve (e.g. the ten-layer case in Kirsch, 2006).

For a macro-anisotropic section, average longitudinal and average transverse resistivities can be calculated from the parameters of the single beds (Kirsch, 2006):

$$\rho_L = \frac{\sum h_i^*}{\sum \frac{h_i}{\rho_i}} \quad (5.45)$$

$$\rho_T = \frac{\sum h_i^* \rho_i}{\sum h_i} \quad (5.46)$$

Substitute resistivity and thickness as well as coefficient of anisotropy are the same as for micro-anisotropic behaviour (eqs. 5.43 and 5.44).

Equivalence between true resistivity sections and sections with substitute layers due to anisotropy may cause serious modelling and interpretation errors if the anisotropy (both macro and micro) is not recognized. As can be calculated from eqs. 5.43–5.46, intermittently occurring high- and low-resistivity beds may lead to large coefficients of anisotropy and, correspondingly, to large errors in modelled thicknesses.

Anisotropy also leads to discrepancies between results of vertical electrical soundings and electromagnetic induction measurements (horizontal current flow lines). Comparing VES data with data from resistivity borehole logging (mostly horizontal flow lines), anisotropy must also be taken into consideration.

5.14 2-D AND 3-D ELECTRIC IMAGING

In many cases, resistivity can change as both depth and horizontal distance increase. Both VES and electric profiling are limited to surveying in one direction. Electric imaging is able to survey both vertical and horizontal changes in resistivity; field layout is shown in www.agiusa.com/agi2dimg.shtml. This method essentially combines the other two methods. Following on the 1-D applications of resistivity imaging theory come the 2-D and, subsequently, 3-D applications (Loke, 2015). The most common configuration of the 2-D survey employs dipole–dipole electrode configurations. An example of the data acquisition geometry for a 2-D profile is presented in www.agiusa.com/agi2dimg.shtml. It shows a transmitting current dipole I followed by a series of potential dipoles V which measure the resulting voltage gradient at each station along the line. Subsequent measurements are completed by sequentially moving the current dipole down the line. However, alternative resistivity measurements can be made using towed surface or marine arrays, which would maintain the preceding configuration and build up the 2-D image by moving the entire measurement array for each series of measurements.

In both cases, the resulting image plots the apparent resistivity with depth, which is then contoured (commonly krigged) using a commercially available program. The colour-contoured image displays the distribution of apparent resistivity values and associated gradients within the area of interest. In order to convert the apparent resistivity data to true resistivity, the data are inverted. Figure 5.27 displays an example of a measured apparent resistivity pseudosection at the top, followed by a calculated apparent resistivity pseudosection, and resulting in the inverted true resistivity 2-D section. The numbers presented at the bottom of the inverted section display goodness-of-fit criteria used to assess the accuracy of the calculated resistivity model. Finally, note that the surface elevations have been included in the final model, accounting for variations in measurement geometry due to changing topography.

5.15 2-D AND 3-D ELECTRICAL RESISTIVITY TOMOGRAPHY

The resolution of electrical imaging rapidly declines with distance from the electrodes. However, the use of cross-hole survey by positioning electrodes closer to the area of interest, and the use of other available data from geotechnical survey, borehole log, etc., can constrain the inversion model to reduce this limitation (Loke et al., 2013). ERT has been extensively used in geotechnical, engineering and environmental (Grellier et al., 2008; Maslakowski et al., 2014), agricultural (Petersen and al Hagry, 2009), archaeological (Xu and Noel, 1993), soil water content (Cosentini et al., 2011), saturation (Abu-Hassanein et al., 1996), compaction (Seladji et al., 2010), pollution (Godio and

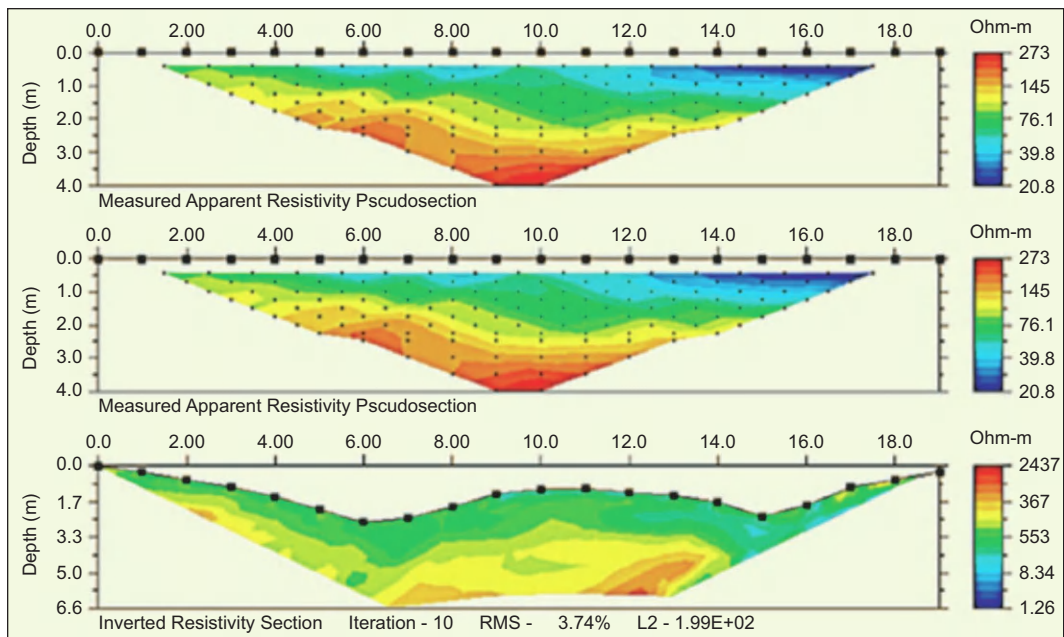


FIGURE 5.27 Examples of measured apparent resistivity, calculated apparent resistivity, and inverted resistivity sections.

Source: www.agiusa.com/agi2dimg.shtml.

Naldi, 2003), mineral exploration (Bauman, 2005; Legault et al., 2008), structures from millimetre scale to kilometres (Storz et al., 2000; Tabbagh et al., 2000; Lebourg et al., 2005; Linderholm et al., 2008), cross-boreholes (Chambers et al., 2003; Daily and Owen, 1991), and similar investigations. Besides, ERT has been adopted as a new technology for long-term embankment monitoring systems through time-lapse and real-time measurements (Barker and Moore, 1998; Gunn et al., 2010; Chambers et al., 2011a, 2011b; Gunn et al., 2014; Merritt, 2014). Also, resolving complex geological problems, such as hidden underground structures identification or studying the spatiotemporal evolution of groundwater flow relative to landslide phenomena accurately, has become easier through its innovative and robust inversion techniques (Dahlin and Zhou, 2006; Loke and Barker, 1996).

5.15.1 ELECTRICAL RESISTIVITY TOMOGRAPHY FIELD SURVEY DESIGN AND DATA ACQUISITION

Some of the resistivity acquisition systems in use are the LUND Resistivity Imaging System (ABEM), MacOhm 21 (DAP-21) Imaging System (OYO), and Sting/Swift, among others. The whole field data acquisition procedure is controlled through the computerized in-built system in any of these sophisticated acquisition systems. The imaging acquisition systems work as both transmitter and receiver and consist of a terrameter (e.g. ABEM Terrameter SAS 1000/4000); electrode selector, also known as the switching unit (e.g. ES 464, ES 10–64C, etc.); multicore cables, usually with a quantity of cable joints (take-outs); a power source; and stainless steel electrodes, to minimize the effects of electrode polarization.

ERT imaging is performed by matching the measured apparent resistivity pseudosection to a computed pseudosection that is obtained by solving for a given Earth resistivity structure $\rho(r)$ using the scaled-Laplace equation (Everett, 2013):

$$\nabla \cdot \left(\frac{1}{\rho} \right) \nabla \varphi = 0$$

The electric potential distribution $\varphi(r)$ is evaluated at the locations of the potential electrodes and transformed into a computed apparent resistivity. The model is then adjusted, and the apparent resistivity recomputed, until it matches the measured apparent resistivity to within a predefined acceptable tolerance (Loke, 2002, 2004; Everett, 2013). ERT surveys can be performed in either 2-D or 3-D, depending on the nature of investigation and the parameter of interest, to be determined in the subsurface. 2-D ERT surveys are carried out using two or four sets of multicore cable reels with a series of take-outs of equal intervals for grounded electrodes to be connected to them via cable jumpers. The length of the multicore cable determines the length of the initial profile without roll-along. The acquisition setups for 2-D ERT measurements (Loke et al., 2013) are in stages after all take-outs have been connected to grounded electrodes. Firstly, the extended connector sockets from the reference electrode are connected to the switching unit. Thereafter, the switching unit is connected to the resistivity meter by a special cable. The resistivity meter is then connected to a power source, usually a car battery. The sequence of electrical measurements, array type, and amount of current to be injected is determined by the resistivity meter, while the switching unit controls which electrodes inject current and which electrodes measure the potential difference. The technique requires the collection of data at several multiples of a (commonly up to $16a$) to provide information at a range of depths, termed n levels. Each n level effectively corresponds to a constant separation traverse at a fixed multiple of a . The acquisition system can be set to run automatically through the required number of levels and also perform noise checking and reacquiring of bad data points. Reciprocal measurements are usually taken to ascertain data quality using a reciprocal error of 5% or 10% as an arbitrary cut-off between good and bad data. The resultant measured resistivity values are inverted using a special software to produce an inverse model resistivity section (Dahlin, 1996; Reynolds, 2011). In 3-D surveys, the pole–pole, pole–dipole, and dipole–dipole arrays are often used because they have better resolution at the edges of the survey grid than other arrays (Loke, 2002). The field layout is usually arranged in a square grid with equal electrode spacing along the x and y directions (Akingboye, 2018 Loke and Barker, 1996; Loke, 2002).

However, a rectangular grid with different numbers of electrodes and spacing may be used for elongated bodies. According to Loke (2002), 3-D surveys are performed in a number of ways, which are categories 1, 2, 3, and 4. In category 1, measurements are taken along possible directions with electrodes arranged in a rectangular grid. Category 2 has all electrodes arranged in a rectangular grid and measurements are taken along all the grid lines, but limited measurements are made at an angle to the grid lines. Category 3 is used when a multielectrode system has limited nodes to cover an entire survey area but measurements are only in two directions along grid lines. In category 4, measurements are taken only along a series of parallel 2-D survey lines, while the 3-D subsurface resistivity model is produced by combining and inverting all the parallel 2-D lines together. However, the distance between parallel 2-D resistivity lines should be equal to the electrode spacing in order to achieve this. For quality data coverage, field measurements should be taken in either category 1 or 2 for angular data to be included in the acquired data, because such information would not be captured in measurements taken in either category 3 or 4.

5.15.1.1 Cross-Hole (Borehole) ERT Survey

The major limitation of ERT survey conducted at the Earth's surface has to do with the resolution of resistivity images with depth because the electrodes do not have direct contact with subsurface layers. Therefore, cross-hole ERT survey is mainly used to improve subsurface image resolution. Cross-hole ERT employs various quadrupole combinations of current and potential electrodes, either in the same hole, between the hole and the surface, or between pairs of holes (Daily et al., 2000; Loke et al., 2013). Loke et al. (2013) present a 2-D electrical resistivity image of the subsurface. In this scenario, a series of electrodes are placed at equivalent intervals vertically down two well casings. Each available dipole is used for both transmitting (current) and receiving (voltage). The separation of the holes should not be more than about 0.75 times the borehole array length in order to achieve an acceptable image resolution (LaBrecque et al., 1996).

5.15.1.2 Mobile Land and Underwater ERT Survey

The mobile ERT imaging system is used on land (e.g. Aarhus Pulled Array System) and underwater (Sorensen, 1996; Bernstone and Dahlin, 1999; Loke, 2002). The underwater mobile ERT imaging technique can be conducted at the water surface with floating electrodes or submerged at the floor to investigate the subsurface condition beneath the river/stream/lake/seafloor through continuous resistivity profiling. The method may be more beneficial to use fixed submerged arrays in extremely shallow streams instead of continuous resistivity profiling to prevent cable damage (Loke et al., 2013). However, floating electrodes are best recommended when a water column is no greater than 25% of the total depth of investigation (Loke and Lane, 2004). The figure in Akingboye (2018; Loke, 2002) shows one of the possible arrangements of underwater mobile imaging systems that employed the Wenner–Schlumberger configuration using a cable with a number of nodes being pulled along the river/lake/sea bottom by a boat. The nodes consist of two fixed-current electrodes ($C1$ and $C2$), while the rest are used as potential electrodes (P) to measure the potential at different spacing. This type of imaging system can be used with other arrays, like the gradient array.

5.15.2 ELECTRICAL RESISTIVITY TOMOGRAPHY INVERSION

The ERT inversion program is a computer program that automatically generates a 2-D resistivity model of the subsurface for a dataset from resistivity imaging surveys of conventional or nonconventional arrays, with an almost-unlimited number of possible electrode configurations of uniform or nonuniform electrode spacing, and for underwater and cross-borehole surveys (Griffiths and Barker, 1993; Dahlin, 1996). The inversion routine used by the program is based on the smoothness-constrained least squares method that supports both the quasi-Newton and the Gauss–Newton least squares optimization methods. The Gauss–Newton method gives slightly better results for a model with large resistivity contrasts greater than 10:1 that can have an erratic resistivity distribution with spurious high- or low-resistivity zones, but slower than the quasi-Newton method (Loke and Barker, 1996; Loke and Dahlin, 2002; Loke, 2002). The smoothness-constrained least squares method is based on

$$\left(J^T J + uF \right) d = J^T g$$

where $F = f_x f_x^T + f_z f_z^T$, f_z is the vertical flatness filter, J is the matrix of partial derivatives, u is the damping factor, d is the model perturbation vector, and g is the discrepancy vector.

This mathematical inverse problem determines the subsurface distribution of resistivity from measurements of apparent resistivity datasets to produce a subsurface inverted model that agrees mostly with the field apparent resistivity measurements based on predefined numbers of iterations for convergence. The subsurface is divided into a lot of rectangular cells during modelling, but the resistivities of the cells are determined by the inversion algorithm but might not always give the ideal resistivities because the cell-based inversion may employ a lot of assumptions to model complex geological structures (Loke, 2002). A summary of the stages involved in inversion processes for data reductions and for generating the final inversion model resistivity section is given in what follows (Akingboye, 2018).

1. Files in the Terrameter are saved in binary formats. This format is not compatible with the inversion software program, so it has to be converted to a software-readable file format.
2. Data editing is performed to remove bad data points that have resistivity values which are clearly wrong due to the failure of the relays at one of the electrodes, poor electrode ground contact due to dry soil, or shorting across the cables due to wet ground conditions.
3. Reciprocal measurement for data assessment and editing may be employed to determine the percentage error resulting from the interchanging of current and potential electrodes

- in contrast to the normal array of two-current and two-potential electrodes (Zhou and Dahlin, 2003; Wilkinson et al., 2012).
4. Splicing is adopted for too large datasets in order to choose a section from the segmented datasets to be processed at a single time (Loke, 2004).
 5. Forward modelling subroutine—finite-difference and finite-element methods. 2-D/3-D model apparent resistivity and Jacobian matrix values are calculated through a mathematical link between the model parameter and the response model provided by finite-difference (Dey and Morrison, 1979a, 1979b; Loke, 1994) or finite-element methods (Silvester and Ferrari, 1990).
 6. The ratio of the vertical flatness filter $\left(\frac{AB}{2} = 306\right)$ to the horizontal flatness filter (f_x) is used for smoothing and reducing elongated vertical and horizontal anomalies in the pseudosection. Higher vertical to horizontal flatness ratio values for elongated vertical anomalies and lower ratio values for elongated horizontal anomalies are considered to produce a better pseudosection (Loke, 2004).
 7. Models with very noisy and less noisy datasets, as well as unnatural oscillations in the lower section, are taken care of during inversion by adjusting the damping factor (u) value.
 8. A relatively large damping factor is used for very noisy datasets and unnatural oscillations, while a smaller damping factor is used for less-noisy datasets to generate a better inversion model. However, a minimum limit of about one-fifth of the initial damping factor for damping factor value may be used to stabilize the inversion process (deGroot-Hedlin and Constable, 1990; Loke, 2004).
 9. The inversion/model refinement allows the user to choose a model with cell width with normal electrode spacing or cell width with half-electrode spacing for optimum result. Cell width with half-electrode spacing gives better resolution and smoothens the model than does cell width of normal electrode spacing, because the program attempts to reduce misfit of failure from the near-surface variations that could lead to distortions in the lower section of the model (Loke, 2004).
 10. For achieving root mean square (RMS) or absolute (Abs.) error less than 10%, the iteration subroutine is usually set to five iterations for convergence but can be continued to about ten iterations in case model RMS/Abs error is higher than the required convergence limit. According to Loke (2004), a model with the lowest RMS error is not always the best model, particularly for very noisy datasets. The figure in US EPA (2019) shows an example of an inverted 2-D cross-borehole electrical resistivity tomography (ERT) dataset.

5.16 APPLICATIONS OF RESISTIVITY SURVEY AND CASE STUDY

5.16.1 IN GROUNDWATER STUDIES

In groundwater studies, the resistivity method can furnish information on subsurface geology which might be unattainable by other geophysical methods. For example, electrical methods are unique in furnishing information concerning the depth of the fresh water–salt water interface, whereas neither gravity, magnetic, nor seismic methods can supply such information. A thick clay layer separating two aquifers usually can be detected easily on a sounding curve, but the same clay bed may be a low-velocity layer in seismic refraction surveys and cause erroneous depth estimates (Zohdy, 1969a). From 1965 to 1969, the US Geological Survey made several resistivity surveys in the South-Western United States, where fresh water–salt water interfaces were mapped successfully with Schlumberger and equatorial electrical soundings. The apparent resistivity map (Zohdy et al.,

1964, modified after Zohdy et al., 1969) was obtained with $\frac{AB}{2} = 306$ m in the White Sands Missile Range area (Zohdy et al., 1969). The apparent resistivity contour of 10 Ω -m delineates, qualitatively, the area where mineralized groundwater is to be expected at shallow depth. The mixing of fresh

and brackish groundwater has been the main issue of environmental problems. In order to facilitate proper management of groundwater resources in the megacity areas, systematic resistivity surveys were carried out in the metropolitan suburbs of Calcutta, West Bengal (Choudhury et al., 1997). The apparent resistivity contour of 10–30 Ω -m delineates, qualitatively, the area where mineralized groundwater is to be expected at shallow depth (Choudhury et al., 1997).

5.16.2 DELINEATION OF SHALLOW CRUSTAL STRUCTURE

Combined Schlumberger and equatorial direct current (DC) dipole soundings were conducted along the Sanwer–Borgaon traverse across the Narmada–Son lineament (NSL) primarily to locate the shallow-level faults and fractures after two intraplate earthquakes that rocked Jabalpur (1997) and Pandhana (1998) in Madhya Pradesh, Central India (Naskar et al., 2002). Interpretation of resistivity data reveals that the thickness of the Deccan trap varies from a few tens of metres to more than 1 km. Bundelkhand granite marks the high resistivity basement. Formations of intermediate to low resistivity are the Gondwana, Vindhyan, Bijawar, and Quaternary sediments (Figure 5.28). There is the presence of a high-resistivity (>1,000 Ω -m) formation at a relatively shallow depth (600–700 m) south of the Narmada River, whereas the same high-resistivity formation occurs at a greater depth to the north of the river. But the basement has come up to a level of 61 m below ground level at the Narmada River. The steep gradient of the basement could possibly be attributed to a fault/rift zone. Therefore, the Narmada–Son lineament is a faulted zone. The Narmada River flows right over the fault.

5.16.3 DELINEATION OF GONDWANA SEDIMENTS BELOW THE DECCAN TRAP

Identification and delineation of the hidden sediments under the cover of the Deccan trap are a difficult problem in geoscientific exploration. The exploration for additional coal measures in the known Gondwana basins and their covered extensions needs immediate attention to supplement

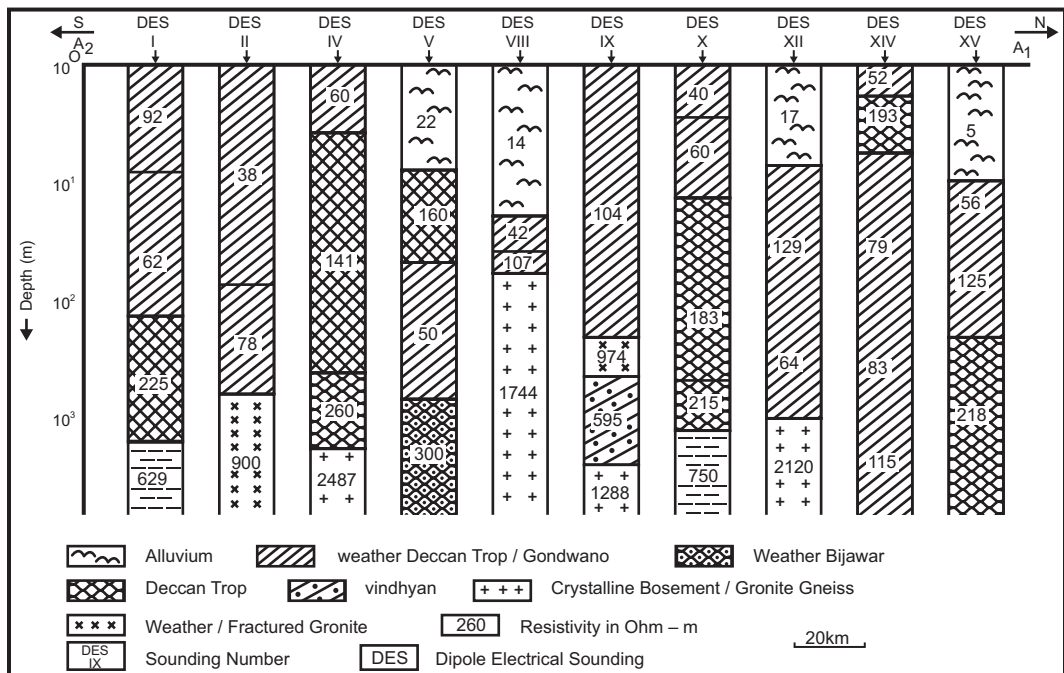


FIGURE 5.28 Interpreted vertical resistivity distribution and inferred lithology across NSL, Central India, MP.

the energy basket. In this context, the delineation of Gondwana sediments below the Deccan trap assumes greater significance. To achieve these objectives, certain areas of the south-western part of the Wardha Valley, Maharashtra (Naskar and Saha, 2015), have been selected for electrical survey employing Schlumberger sounding. The result of resistivity sounding (RS) data interpretation shows mainly four interfaces: topsoil/alluvium, Deccan trap, Gondwana sediment, and basement of high resistivity. Along the section AA', the topmost layer is characterized by resistivity value of 6–15 Ω -m and thickness varying from 2 to 15 m corresponding to the alluvium/topsoil, which is underlain by a layer having resistivity varying from 25 to 325 Ω -m and of thickness varying from 50 to 80 m. This layer corresponds to the Deccan trap. These are followed by a low resistivity zone of 11–104 Ω -m and of thickness 110–205 m, representing subtrappean formations Motur/Barakar. A thick conductive zone of resistivity varying from 6 to 18 Ω -m and thickness varying from 340 to 700 m representing Talchir is interpreted all along the section. The high-resistivity zone (69–500 Ω -m) representing the basement is inferred at a deeper level (Naskar and Saha, 2015). The results of resistivity soundings are in agreement with the gravity results along the section. Basement depression is found to lie in the central portion of the Bouguer anomaly map (Joga Rao et al., 1974, 1984). The geoelectric section indicates a fault zone in the Mangli–Sindhi segment due to an appreciable large variation in basement depth (871 m and 313 m) in two adjoining RS locations.

5.16.4 DETECTION OF SUBSURFACE VOIDS

In order to restore environmental equilibrium in the affected areas (land subsidence over abandoned underground mines), the pole–dipole resistivity technique has been employed over an area of 0.18 km² in the Baktarnagar area, adjacent to Ghanashyam Colliery, Raniganj Coalfield area, Burdwan District, West Bengal (Biswas et al., 2005). The resistivity method, in conjunction with Bristow's graphical interpretation procedure, could delineate a number of air-filled and solution-filled subsurface cavities based on high- and low-resistivity perturbations, between 21 m and 70 m depth in the area (Biswas et al., 2005). The case study described reveals that, under favourable conditions, guide the location of boreholes for stowing and stabilizing the old workings.

5.16.5 DELINEATION OF COAL-BEARING GONDWANA ROCKS

A geophysical survey employing Schlumberger resistivity technique has been carried out in order to delineate the basement topography beneath the outcropping coal-bearing Gondwana rocks over a virgin Gondwana prospect in the vicinity of Bishrampur and the Tatapani–Ramkhola coalfield area, Surguja District, Chhatisgarh (Naskar et al., 2008). Deep resistivity sounding confirm that the overlying Gondwana rocks represent the Barakar with a resistivity range of 50–400 Ω -m and thickness varying from 1 to 442 m. The Archaean basement is characterized by a higher order of resistivity (Naskar et al., 2008).

5.16.6 3-D CONFIGURATION OF KIMBERLITE BODIES

A resistivity method employing the Schlumberger sounding technique with $\frac{AB}{2}$ as large as 2 km has been conducted in order to map the eastern and northern extensions of an already-known kimberlite sill in a grid pattern of 200 m $\times$ 200 m over a 1 km² area between the Tokapal and Duganpal Villages on the Jagdalpur–Gidam road about 18–20 km from Jagdalpur Township, Chhatisgarh. Kimberlite is exposed in a pit and intersected in some drill holes at a shallow depth to the west of the area surveyed (Das et al., 2005). Because of overlapping resistivity of the metasediments (100–500 Ω -m) and the basement rock, it has not been possible to map the basement everywhere. Even the bottom of the kimberlite has been eluding in some sounding curves because of an overlapping order of resistivity of the kimberlites in more massive form with that of the metasediments in the area.

Five of these soundings have been tied with drilling results already available in the area for proper correlation of the resistivity parameters with the actual geometry of the kimberlite as met in the boreholes. These data have been used to prepare a subsurface 3-D disposition of the kimberlite in the area using RockWorks. This is presented in Das et al. (2005). Just to elucidate the same, one 2-D E–W geological section is also shown in Das et al. (2005) along the station N200 from W400 to E400.

5.16.7 DEEP CRUSTAL STUDIES

To delineate the thicknesses of the lower Siwaliks/Gondwana formations above the basement, geophysical survey was carried out using deep electrical resistivity techniques along the Purnea-Dalkhola-Sevoke section. Deep electrical surveys were undertaken using azimuthal dipole–dipole sounding with R (separation between transmitting and receiving dipoles) as large as 8–10 km (Naskar, 2016). The shallower part up to $\frac{AB}{2} = 3$ km was covered by using Schlumberger sounding to avoid the effects of lateral inhomogeneity. A notable feature is observed in between Siliguri and Sevoke (ES-14), where a resistivity value of 1,100 Ω -m near the surface persists with depth. However, after 40 m depth, this resistivity value suddenly falls. This is interpreted to be due to the presence of Siwaliks at depth. The high-resistivity basement is encountered at a depth varying from 2,917 to 4,450 m around the Dalkhola and Kishanganj area s(ES-1 and 5). An ONGC borehole (500 m from ES-2) towards Purnea encountered Siwaliks (around 2 km), followed by upper (369 m) and lower (1253 m) Gondwana over the Archaean basement. Some distinct breaks (faults) are also observed. These faults are located near km st 452, 493, 510, and north of km st 526, respectively (Figure 5.29). The relative displacements of the basement are shown by arrows depicted on the figure around the interpreted faults (Naskar, 2016). Of particular interest is the zone between km st 452 and 493, where resistivities to the tune of 1,000–2,000 Ω -m are mapped at comparatively shallower

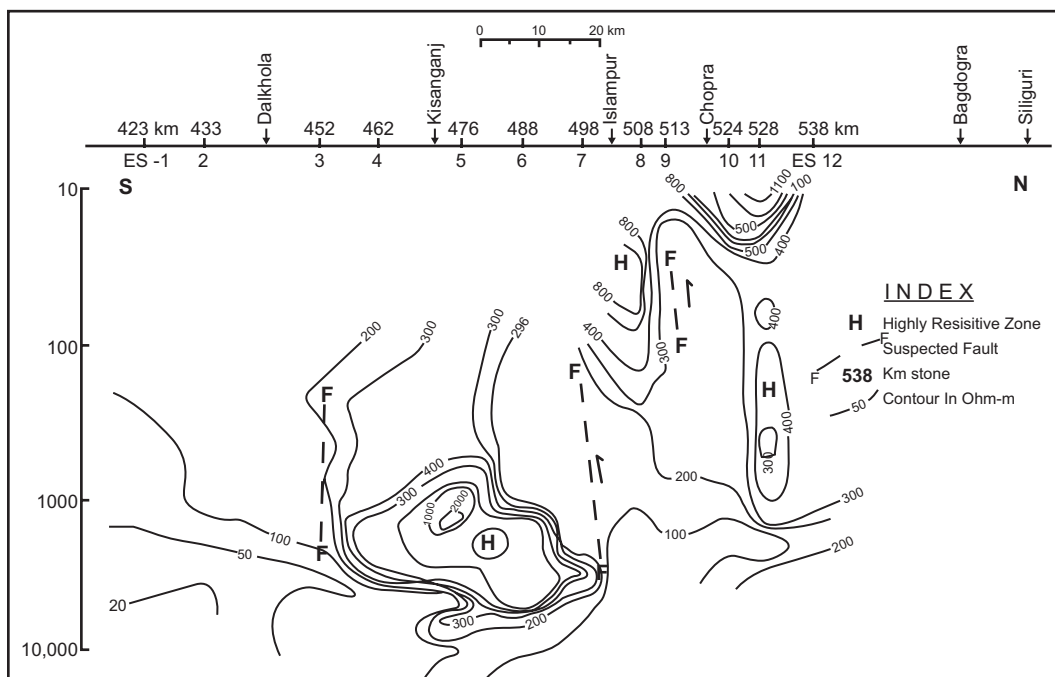


FIGURE 5.29 Apparent resistivity pseudodepth section along Purnea-Dalkhola-Sevoke.

level. This is interpreted as the basement resistivity, and the basement is up-arched (convex upward) in this zone. This zone may be interesting for Gondwana sediments at shallower depth and/or oil and gas. In the rest of the area north of km st 493, the basement deepens further, with another fault at km st 524.

5.16.8 GEOPHYSICAL IMAGING OF ALPINE ROCK GLACIERS: MURTÈL CASE STUDY (TOMOGRAPHIC GEOELECTRIC SURVEY)

The lightweight equipment required to perform geoelectric surveys can be transported to areas with difficult access. The application of tomographic inversion techniques to geoelectric data requires a large number of measurements, which can be made efficiently with modern multichannel systems that include several tens of electrodes (Stummer et al., 2002). The most significant problem in geoelectric surveying on rock glaciers is the weak galvanic coupling associated with the generally high contact resistances. To compensate for this, water-soaked sponges to the electrodes and water to the immediate vicinity of the electrodes is attached (Maurer and Hauck, 2017). Nevertheless, the currents injected into the ground and the measured voltages are generally low. This issue needs to be considered when selecting the electrode configuration to be used. Because electrode configurations with low geometrical factors provide larger voltages (Reynolds, 1997), Wenner configurations are generally more suitable for surveying rock glaciers than dipole–dipole configurations. Since snow cover may prohibit the current injection, it is preferred to conduct geoelectric surveys in the summer.

Over the past two decades, tomographic inversion techniques for geoelectric data have become increasingly popular and user-friendly. The application of these programs to mountain permafrost data poses no serious problems so long as they correctly account for the effects of the pronounced topographic relief.

Maurer and Hauck (2017) conducted 103 Wenner geoelectric measurements along a 145 m long profile across the frontal region of the rock glacier, deploying 30 electrodes equally spaced at 5 m intervals. Electrode coupling was especially difficult on this rock glacier, because the surface consisted of large boulders (up to 2 m high) with only small amounts of fine-grained material. The total acquisition time was 5 hours. The geoelectric data were tomographically inverted using RES2DINV, a commercial software package (www.abem.se/). A moderately conductive near-surface layer (DC1) with a resistivity of ~ 10 k Ω -m and a thickness of about ± 2.5 m is observed along the length of the recording line (Maurer and Hauck, 2017).

Within the rock glacier, that is, below DC1, in the horizontal distance range 50–140 m, the resistivity increases abruptly to the M Ω -m range. At about 140 m horizontal distance, near where the surface exposure of the rock glacier ends, resistivities at depth change from the M Ω -m range to the k Ω -m range across a steeply dipping boundary denoted as DC2 *in*. The very high resistivities in Maurer and Hauck (2017) are evidence for high ice content within the core of the rock glacier. This is consistent with the borehole log. A sharp lateral change to lower resistivities at DC2 delineates the beginning of unfrozen material at the front of the rock glacier, and the isolated high-speed body, S3, probably represents a large boulder (Maurer and Hauck, 2017) not represented in the other geophysical datasets. The seismic speed discontinuity, S2, in Maurer and Hauck (2017) defines the top of the bedrock. Towards the frontal part of the rock glacier, S2 becomes shallower, reaching the surface at about 170 m horizontal distance. The high bedrock speeds within this bedrock protrusion may indicate the presence of virtually unfractured rock that may have acted as a barrier to the flow of the rock glacier.

5.16.9 GEOPHYSICAL IMAGING OF ALPINE ROCK GLACIERS: MURAGL CASE STUDY (TOMOGRAPHIC GEOELECTRIC SURVEY)

Maurer and Hauck (2017) employed very similar recording parameters and identical data inversion software for the Murtèl and Muragl rock glacier investigations. The resultant uninterpreted and interpreted resistivity tomograms for the Muragl rock glacier are shown, respectively. A thin

layer of moderately high-resistivity material underlies the undulating surface (Maurer and Hauck, 2017). This is underlain in the south-western part of the survey line by a higher-resistivity material (Maurer and Hauck, 2017). Within *DC2*, there is a highly resistive core with resistivities in the $M\Omega$ -m range (Maurer and Hauck, 2017). The remaining regions of the tomogram have resistivities of several $k\Omega$ -m. The active layer is delineated by the geoelectric and seismic layers *DC1* and *S1* (Maurer and Hauck, 2017). Since snow covered the ground during the seismic survey, *S1* overestimated the thickness of the active layer.

The presence of ice is identified in the south-western part of the rock glacier by a combination of high resistivities and seismic speeds of about $3,500 \text{ ms}^{-1}$ (Maurer and Hauck, 2017). The resistivities are generally lower than those observed within the Murtèl rock glacier, perhaps indicating a higher proportion of unfrozen water and related warmer temperatures. Resistivities in the $M\Omega$ -m range are only detected within the relatively small region, *DC3*. Although it is difficult to compare results from the longitudinal Murtèl profile with the perpendicular Muragl profile, based on the distributions of resistivity in Maurer and Hauck (2017) and Maurer and Hauck (2017), it seems likely that the distribution of ice in the Muragl rock glacier is rather more heterogeneous than in the Murtèl rock glacier. The low-seismic-speed block, *S2*, is distinguished by moderate resistivities, which, according to borehole log *BI*, are due to the presence of voids filled with air (low seismic speed) and minor occurrences of water (decreased resistivities). Results from surface displacement analyses (Kaab and Vollmer, 2000) indicate that movements have practically ceased within this region. Taken together, these observations indicate that the permafrost has degraded in this part of the rock glacier (i.e. the ice has melted, leaving behind large voids).

5.16.10 THE USE OF RESISTIVITY METHOD IN THE INTERNAL CHARACTERIZATION OF LANDFILLS, ILHAVO, PORTUGAL: A CASE STUDY

As contamination is time- and space-dependant, 4-D studies have been carried out, that is, measurements have been repeated in the same locations but at different times (Brewster et al., 1995; Matias and Hermozilha, 2006). More recently, the internal structure of landfills has also been studied with success by various geophysical methods (Biandini et al., 2004; Bernstone et al., 2000; Soupios et al., 2007). Because of the complexity of the problems, integrated approaches have been favoured (Grenn et al., 1999) as they allow a cross-correlation of interpretations from the methods used. The most frequent methods are resistivity, electromagnetics, ground-probing radar, and seismics.

Three resistivity 2-D sections were carried out on the landfill using the Wenner/Schlumberger array, RP3, RP5, and RP8 in Mantlik et al. (2009). The RP8 section runs more or less parallel to the gravity profiles, while RP3 and RP5 are roughly perpendicular to the gravity profiles. The aim of these 2-D sections is to give a priori information to be used in gravity data modelling. It is therefore intended to estimate the thickness of the landfill deposits, its lateral extension, as well as the eventual lateral changes in the internal landfill composition. Resistivity modelling was done using the RES2Dinv software and assisted with information from nearby boreholes *S1/S2*, *S4*, *S9/S10* (Mantlik et al., 2009). Resistivity data are obtained over very heterogeneous data, but nevertheless, rms errors of 9.8%, 7.6%, and 8% were obtained for profiles RP3, RP5, and RP8, respectively, after ten iterations. The interpretation of the profile RP8 is depicted in Mantlik et al. (2009).

As shown, there is some lateral change of the resistivity in the extent of the profile. The landfill deposits edge is clearly limited at position 30 m on the profile. The landfill deposits are delimited into three zones of low resistivity less than $10 \Omega\text{m}$, divided by narrow zones of higher resistivity at 100 and 160 m. The landfill bottom is estimated to be at depths varying from 8 to 10 m, represented by the depth of the significant resistivity increase along the profile. The interpretation of the profile RP3 is shown in Mantlik et al. (2009), and for the profile RP5 in Mantlik et al. (2009). These two resistivity profiles are perpendicular to the gravity profiles (Mantlik et al., 2009).

The lateral boundary of the landfill is well marked at position 170 m in both profiles. In both profiles, significant resistivity inhomogeneities are evident. In profile RP3, there are three low-resistivity

zones $<10 \Omega\text{m}$, divided by two resistivity zones at the coordinates 70–80 and 100–110 m. The profile RP5 has a similar internal structure, with a high resistivity zone at positions 65–80 m and 130–140 m. The estimate for the thickness of the landfill at both profiles RP3 and RP5 ranges from 8 to 10 m, as in profile RP8. In depth, the resistivity model allows the following interpretation:

- A topmost resistive layer corresponds to the landfill cover, thickness ranging from 1 to 2 m.
- A second, conductive layer, that is, the signature of the landfill deposits and with a thickness varying from 7 to 8 m. As can be seen in all sections, this second layer is quite inhomogeneous.
- Finally, the bottom, more resistive layer corresponds to the local geology, in this case organic mud.

The resistivity profiles interpretation results constrained the thickness of landfill deposits.

5.16.11 RESISTIVITY METHOD FOR ARCHAEOLOGICAL SITE CHARACTERIZATION OF JORDAN, A COUNTRY IN WESTERN ASIA: CASE STUDY

Galvanic or soil conduction electrical methods have been used since the 1950s (Aitken, 1974). The best-known method is resistivity profiling, and typically a Wenner, dipole–dipole, and pole–pole arrays. Sumner (1976), Batayneh and Al-Zoubi (2000), Batayneh (2001), Batayneh and Al-Diabat (2002), Batayneh and Barjous (2003), and Batayneh (2005) have provided a good description of methods and arrays in common use along with their relative advantages. The electrical 2-D resistivity tomography (Wenner array) results from the Umm er-Rasas archaeological site in Central Jordan have been discussed by Batayneh et al. (2006). The profile shows a conductive zone as a result of water seepage under the architectural features of the site of Umm er-Rasas. The induced polarization (IP) and self-potential (SP) methods are also soil-contacting methods. They were first used in England in 1968 (Aspinall and Lynam, 1968, 1970). The SP method was used for reconnaissance archaeological mapping by Wynn and Sherwood (1984). Because of the too small signals in a field survey caused by oxidized metal burial artefacts, the use of induced polarization technique in archaeology was limited.

5.16.12 APPLICATION OF VERTICAL ELECTRICAL SOUNDING (VES) AND ELECTRICAL RESISTIVITY TOMOGRAPHY (ERT) IN LOCATING FRESH WATER IN COASTAL TERRAIN AT VIZAG STEEL PLANT (VSP), VISAKHAPATNAM, ANDHRA PRADESH, INDIA: CASE STUDY

The entire area was underlain by recent alluvial of varying thickness consisting of sand, clay, and silt, with basement as Khondalites rock formation consisting of saline water in some parts of the area. Locating favourable groundwater locations in such a terrain is a difficult task. The area consists of saturated clay, saturated silt, and saline water with the same resistivity signature and misleads resistivity data interpretation for identifying and recommending the suitable groundwater locations in a coastal terrain. To overcome these uncertainties, a second geophysical approach or combinations of geophysical methods become necessary to resolve these resistivity ranges in a more reliable and better, fruitful interpretation of subsurface layers. Vertical electrical sounding (VES) (Stampolidis et al., 2005; Kumar et al., 2007; Yadav and Singh, 2007; Dhakate et al., 2008, 2012) and electrical resistivity tomography (ERT) (Olorunniwo and Olorunfemi, 1987; Olayinka and Mbachi, 1992; Emenike, 2001; Ariyo and Banjo, 2008; Gabr, 2011) are the most widely used geophysical techniques for hydrology and groundwater exploration.

In order to assess the groundwater conditions at VSP and township, 94 VESs using Schlumberger configuration with 100–200 m half-current electrode separations were carried out. VES data are

initially interpreted using the curve matching technique (Orellana and Mooney, 1966), and then by inversion iteration method (Jupp and Vozoff, 1975), to arrive at absolute resistivities values of subsurface layers (Dhakate et al., 2015). Similarly, at few selected places, ERTs using the Wenner–Schlumberger configuration with 5 m interelectrode spacing using 48 electrodes were carried out (Dhakate et al., 2015). Based on the interpretation of integrated geophysical approaches, three sites were initially recommended for drilling of borewells (Dhakate et al., 2015). First borewell was recommended and drilled at a full material disclosure (FMD) area up to a depth of 100 m at VES no. 44, whose resistivities are 44, 18, 46, and 569 Ωm , with thicknesses 1, 12, and 17 m, respectively. The resistivity of 18 and 46 Ωm belongs to a highly weathered and saturated weathered formation, followed by a high-resistivity zone beneath. The interpreted VES curve is shown in Dhakate et al. (2015). ERT carried out nearby this location correlates with the VES interpreted data. From the ERT pseudosection (Dhakate et al., 2015) it is observed that a resistivity range of 7 to 14 Ωm occurs around a depth of 13 m, and a resistivity range of 26–48 Ωm occurs around a depth of 30 m, which very well correlates with the VES interpreted results. The ERT pseudosection is shown in Dhakate et al. (2015). A comparison of the observed drilled litholog with interpreted VES results is shown in Dhakate et al. (2015).

The second borewell was also recommended and drilled in the FMD area at VES no. 45, whose resistivities are 33, 17, 36, and 333 Ωm , with thicknesses 1.6, 4.8, and 46 m, respectively. The resistivity of 17 and 36 Ωm represents highly weathered and weathered formations, followed by high resistivity. The interpreted VES curve is shown in Dhakate et al. (2015). ERT carried out at this location also correlates with VES data. The ERT pseudosection represented a resistivity range of 4–14 Ωm up to a 10 m depth, indicating a highly weathered formation, followed by resistivity range of 26–48 Ωm up to a depth of 30 m, representing weathered formation; after 30 m, a resistivity range of 80–162 Ωm with a thickness of 6–7 m was observed, which is not encountered in VES interpretation, representing a fracture zone followed by a high-resistivity zone. The ERT pseudosection is shown in Dhakate et al. (2015). A comparison of observed drilled litholog with interpreted VES results is shown in Dhakate et al. (2015).

The third borewell was recommended and drilled at pump house no. 3 at VES no. 85, whose resistivities are 44, 5.4, and 16 Ωm , with thicknesses 0.5 and 40 m, respectively. The resistivity of 5.4 Ωm represents a clayey/saline water formation, followed by a highly weathered formation. The interpreted VES curve is shown in Dhakate et al. (2015). ERT carried out at this location represents a resistivity range of 5–8 Ωm up to a depth of 25 m, indicating a clay/saline nature of the subsurface; another zone of resistivity 14–24 Ωm occurred at a depth from 25 to 40 m, indicating highly weathered formation, followed by a resistivity of 40–120 Ωm representing weathered formation beyond a 40 m depth. A good correlation was observed between VES and ERT at this location. The ERT pseudosection is shown in Dhakate et al. (2015). A comparison of observed drilled litholog with interpreted VES results is shown in Dhakate et al. (2015). From the interpreted resistivity data, the range of 20–80 Ωm was found to be a good zone for groundwater exploration in this area. The recommended drilled borewells were found to yield in the range of 104–3623 gal/h fresh water to meet the demand of the Vizag Steel Plant. The observed drilled lithologs were well correlated with the VES, ERT, and GPR data. Besides, this hydrochemical analysis of water samples was carried out from the existing borewells and dug wells. Total dissolved solids (TDS) and chloride concentrations ranged from 189 to 3,398 mg/l and 5 to 1,610 mg/l. The eastern and north-eastern parts of the area have an elevated concentration of TDS and chloride, indicating intrusion of saline water. TDS concentration of drilled borewell was found to range from 400 to 500 mg/l. The observed yield of drilled borewells ranges from 104 to 3,623 gal/hr (Dhakate et al., 2015).

REFERENCES

- Abu-Hassanein, Z., Benson, C. and Blotz, L., 1996. Electrical resistivity of compacted clays. *Journal of Geotechnical Engineering*, vol. 122, pp. 97–406.

- Aitken, M., 1974. *Physics and Archaeology*. 2nd edn. Oxford: Clarendon Press.
- Akingboye, A.S., 2018. Two-dimensional geoelectrical subsurface classification of tioro-Akoko area, Ondo State, Southwestern Nigeria. Unpublished MSc Dissertation, Department of Earth Sciences, Adekunle Ajasin University, Akungba-Akoko, Ondo State, Nigeria, p. 110.
- Alpin, L.M., 1950. The theory of dipole sounding. In: Alpin, L.M., Berdichevskii, M.N., Verdrainstev, G.A. and Zagarmistr, A.M. (eds) *Dipole Method for Measuring Earth Conductivity*. New York: Consultant Bureau, pp. 1–60.
- Alpin, L.M., 1958. Transformation of sounding curves. In: *Prikladnaya Geofizika*, vol. 19, pp. 23–46. [Translation in Dipole methods for measuring earth conductivity]: New York: Plenum Press, 1966, pp. 61–78.
- Alpin, L.M., Berdichevskii, M.N., Vendrintsev, G.A. and Zagarmistr, A.M., 1966. *Dipole Method for Measuring Earth Conductivity* (translated by G.V. Keller). New York: Consultants Bureau, p. 302.
- Archie, G.E., 1942. The electric resistivity log as an aid in determining some reservoir characteristics. *American Institute of Mining and Metallurgical Engineers Transactions*, vol. 146, pp. 54–62.
- Ariyo, S.O. and Banjo, A.A., 2008. Application of electrical resistivity method for groundwater exploration in a sedimentary terrain: A case study of Ilara-Remo, South-Western Nigeria. *Continental Journal of Earth Sciences*, vol. 3, pp. 53–58.
- Aspinall, A. and Lynam, J., 1968. Induced polarization as a technique for archaeological surveying. *Prospezioni Archeologiche*, vol. 3, pp. 91–93.
- Aspinall, A. and Lynam, J., 1970. An induced polarization instrument for detection of near surface features. *Prospezioni Archeologiche*, vol. 5, pp. 67–76.
- Barker, R. and Moore, J., 1998. The application of time lapse electrical tomography in groundwater studies. *Leading Edge*, vol. 17, pp. 1454–1458.
- Batayneh, A., 2001. Resistivity imaging for near-surface resistive dyke using two-dimensional DC resistivity techniques. *Journal of Applied Geophysics*, vol. 48, pp. 25–32.
- Batayneh, A., 2005. 2D electrical imaging of an LNAPL contamination, Al Amiriyya fuel station. *Jordan Journal of Applied Sciences*, vol. 5, pp. 52–59.
- Batayneh, A. and Al-Diabat, A., 2002. Application of 2-D electrical tomography technique for investigating landslides along AmmanDead Sea highway. *Jordan Environmental Geology*, vol. 42, pp. 399–403.
- Batayneh, A. and Al-Zoubi, A., 2000. Detection of a solution cavity adjacent to a highway in southwest Jordan using electrical resistivity methods. *Journal of Environmental and Engineering Geophysics*, vol. 5, pp. 25–30.
- Batayneh, A. and Barjous, M., 2003. A case study of dipole-dipole resistivity for geotechnical engineering from Ras en Naqab area, south Jordan. *Journal of Environmental and Engineering Geophysics*, vol. 8, pp. 31–38.
- Batayneh, A., Khataibeh, J., AlRshdan, H., Tobasi, U. and Al-Jahed, N., 2006. The use of microgravity, magnetometry and resistivity surveys for the characterization and preservation of an archaeological site of Umm er-Rasas. *Jordan Archaeological Prospection*, vol. 14, pp. 60–70.
- Bauman, P., 2005. 2-D resistivity surveying for hydrocarbons—a primer. CSEG Recorder, pp. 25–33.
- Berdichevskii, M.N. and Petrovskii, A.D., 1956. Methods of bilateral equatorial sounding. *Prikladnaya Geofizika*, vol. 14, pp. 97–114. [Translation by Ivan Mittin, U.S. Geol. Survey Library, Denver, p. 37].
- Berdichevskii, M.N. and Zagarmistr, A.M., 1958. Problems of interpretation of bilateral dipole electrical soundings of dipole arrays. *Prikladnaya Geofizika*, vol. 19, pp. 57–108. [Translation in Dipole methods for measuring earth conductivity: Plenum Press, 1966, pp. 79–113].
- Bernstone, C. and Dahlin, T., 1999. Assessment of two automated electrical resistivity data acquisition systems for landfill location surveys: Two case histories. *Journal of Environmental and Engineering Geophysics*, vol. 4, pp. 113–122.
- Bernstone, C., Dahlin, T., Ohlsson, T. and Hogland, W., 2000. DC resistivity mapping of the internal landfills structure: Two pre-excavation surveys. *Environmental Geology*, vol. 39, pp. 360–371.
- Bewley, L.V. 1963. *Two-Dimensional Fields in Electrical Engineering*. New York: Dover.
- Bhattacharya, P.K. and Patra, H.P., 1968. *Direct Current Geoelectric Sounding Principles and Interpretation*. Amsterdam: Elsevier, p. 139.
- Biandini, G., Santarato, G. and Vaccaro, C., 2004. Geochemical characterization and geophysical mapping of landfills: The Marozzo canal study (NE Italy). *Environmental Geology*, vol. 45, pp. 439–447.
- Biswas, U.K., Ghosh, D.C., Naskar, D.C. and Chakraborty, P.K., 2005. Detection of subsurface voids by surface geophysical methods in Baktarnagar area of Raniganj coalfield, W.B. Proceedings of Workshop on

- Solid Fossil Fuel. An introspection and perception in Indian context, Geol. Surv. Ind. Spl. Publ. no. 81, pp. 198–210.
- Breusse, J.J., 1963. Modern geophysical methods for subsurface water exploration. *Geophysics*, vol. 28, no. 4, pp. 633–657.
- Breusse, J.J. and Astier, J.L., 1961. Etude des diapirs en Alsace et Baden-Wurtemberg par la method du rectangle de resistivite. *Geophysical Prospecting*, vol. 9, pp. 444–458.
- Brewster, M.L., Annan, A.P., Greenhouse, J.P., Kueper, B.H., Olhoeft, G.R., Redman, J.D. and Sander, K.A., 1995. Observed migration of a controlled DNAPL release by geophysical methods. *Groundwater*, vol. 33, pp. 977–987.
- Burger, H.R., 1992. *Exploration Geophysics of the Shallow Subsurface*. Upper Saddle River, NJ: University of Michigan; Prentice Hall, Inc.
- Chambers, J.E., Gunn, D.A., Meldrum, P.I., Wilkinson, P.B., Ogilvy, R.D., Haslam, E., Holyoake, S. and Wragg, J., 2011a. Volumetric imaging of earth embankment internal structure and moisture movement as a tool for condition monitoring. 11th International Conference and Exhibition, University of Westminster, London, UK, June 29–30.
- Chambers, J.E., Loke, M.H., Ogilvy, R.D. and Meldrum, P.I., 2003. Non-invasive monitoring of DNAPL migration through a saturated porous medium using electrical impedance tomography. *Journal of Contaminant Hydrology*, vol. 68, pp. 1–22.
- Chambers, J.E., Wilkinson, P.B., Kuras, O., Ford, J.R., Gunn, D.A., Meldrum, P.I., Pennington, C.V.L., Weller, A.L., Hobbs, P.R.N. and Ogilvy, R.D., 2011b. Three dimensional geophysical anatomy of an active landslide in Lias Group mudrocks, Cleveland Basin, UK. *Geomorphology*, vol. 125, pp. 472–484.
- Choudhury, K., Saha, D.K., Rai, M.K. and Naskar, D.C., 1997. Geophysical surveys for hydrogeological and environmental studies in the Eastern Calcutta Metropolis. *Indian Minerals (A Journal of Earth Sciences from Geological Survey of India)*, vol. 51, nos. 1–2, pp. 41–56.
- Compagnie GBnerale de Geophysique, 1963. *Master curves for Electrical Sounding*. 2nd revised edn. Leiden: Eur. Assoc. of Explor. Geophys.
- Cosentini, R.M., Vecchia, G.D., Foti, S., and Musso, G., 2011. Detecting water content changes in sand samples by means of Electrical Resistivity Tomography. In: Alonso & Gens (eds), *Unsaturated Soils*. London: Taylor and Francis Group.
- Dahlin, T., 1996. 2-D resistivity surveying for environmental and engineering applications. *First Break*, vol. 14, pp. 275–284.
- Dahlin, T. and Zhou, B., 2006. Multiple-gradient array measurements for multichannel 2D resistivity imaging. *Near Surface Geophysics*, pp. 113–123.
- Daily, W. and Owen, E., 1991. Cross-borehole resistivity tomography. *Geophysics*, vol. 56, pp. 1228–1235.
- Daily, W., Ramirez, A., Binley, A. and LaBrecque, D., 2000. Electrical resistance tomography—theory and practice. *The Leading Edge*, pp. 573–598, DOI: 10.2172/15010154.
- Dakhnov, V.N., 1963. *Electrical Prospecting for Petroleum and Natural Gas Deposits*. Moscow: Gostoptekhizdat, p. 497.
- Das, L.K., Das, A.K., Ghatak, T.K., Mall, R.P., De, S.K., Das, H.K., Saha, D.K., Rai, M.K., Naskar, D.C., Mukherjee, K.K., Hasan, K. and Vuba, C., 1999. Some results of deep electrical resistivity survey in India. GSI Spl. Pub. (Golden Jubilee Seminar on Exploration Geophysics in India), no. 49, pp. 113–130.
- Das, L.K., Das, B., Choudhury, S.N., Naskar, D.C., Karunakar, G. and Dey, S.K., 2005: 3-D configuration of Kimberlite bodies, Indravati Basinal Area, Bastar district, Chhattisgarh. *Geological Society of India*, vol. 65, pp. 679–688.
- Das, L.K., Pathak, C.S., Singh, N.P., Saha, D.K., Rai, M.K. and Naskar, D.C., 1997. A look at the metallogeny of Singhbhum, India: A geophysical approach. *Indian Minerals (A Journal of Earth Sciences from Geological Survey of India)*, vol. 51, nos. 1–2, pp. 107–122.
- De Groot-Hedlin, C. and Constable, S.C., 1990. Occam's inversion to generate smooth two-dimensional models from magnetotelluric data. *Geophysics*, vol. 55, pp. 1613–1624.
- Deppermann, K., 1954. Die Abhangigkeit desscheinbaren Widerstandes von Sondenabstand bei der Vierpunkt-Method. *Geophysical Prospecting*, vol. 2, pp. 262–273.
- Dey, A. and Morrison, H.F., 1979a. Resistivity modelling for arbitrarily shaped two-dimensional structures. *Geophysical Prospecting*, vol. 27, pp. 1020–1036.
- Dey, A. and Morrison, H.F., 1979b. Resistivity modelling for arbitrarily shaped three-dimensional structures. *Geophysical*, vol. 44, pp. 753–780.

- Dhakate, R., Chowdhary, D.K., Gurunadha Rao, V.V.S., Tiwary, R.K. and Sinha, A., 2012. Geophysical and geomorphological approach for locating groundwater potential zones in Sukinda chromite mining area. *Environmental Earth Sciences*, vol. 66, no. 8, pp. 2311–2325.
- Dhakate, R., Kumar, V.S., Amarender, B., Sankaran, S. and Rao, V.V.S.G., 2015. Integrated geophysical and hydrochemical approach for locating fresh water locations in a coastal terrain. *Water Resources Management*, vol. 29, pp. 3401–3417.
- Dhakate, R., Singh, V.S., Negi, B.C., Chandra, S. and Ananda Rao, V., 2008. Geomorphological and geophysical approach for locating favorable groundwater zones in granitic terrain, Andhra Pradesh, India. *Journal Environmental Management*, vol. 88, pp. 1373–1383.
- Dobrin, M.B., 1960. *Introduction to Geophysical Prospecting*, 2nd edn. New York: McGraw-Hill Book Co., Inc., p. 446.
- Emenike, E.A., 2001. Geophysical exploration for groundwater in a Sedimentary Environment: A case study from Nanka over Nanka Formation in Anambra Basin, South-eastern Nigeria. *Global Journal of Pure Applied Sciences*, vol. 7, no. 1, pp. 1–11.
- Everett, M.E., 2013. *Near-Surface Applied Geophysics*. Cambridge, UK: Cambridge University Press, Edinburgh Building, p. 441.
- Evjen, H.M., 1938. Depth factors and resolving power of electrical measurements. *Geophysics*, vol. 3, pp. 78–98.
- Gabr, A., 2011. Groundwater investigation using ERT at Al-Ain, UAE: A Case Study. *IUP Journal of Earth Science*, vol. 5, no. 1, pp. 22–36.
- Ghosh, D.P., 1971. Inverse filter coefficient for the computation of apparent resistivity standard curves for a horizontally stratified earth. *Geophysical Prospecting*, vol. 19, no. 4, pp. 769–775.
- Godio, A. and Naldi, M., 2003. Two-dimensional electrical imaging for detection of hydrocarbon contaminants. *Near-Surface Geophysics*, vol. 13, pp. 131–137.
- Grellier, S., Guerin, R., Robain, H., Bobachev, A., Vermeersch, F. and Tabbagh, A., 2008. Monitoring of leachate recirculation in a bioreactor landfill by 2-D electrical resistivity imaging. *Journal of Environmental and Engineering Geophysics*, vol. 13, no. 4, pp. 351–359.
- Grenn, A., Lanz, E. and Maurer, H., 1999. A template for geophysical investigations of small landfills. *Leading Edge*, vol. 18, pp. 248–254.
- Griffiths, D.H. and Barker, R.D., 1993. Two-dimensional resistivity imaging and modelling in areas of complex geology. *Journal of Applied Geophysics*, vol. 29, pp. 211–226.
- Gunn, D.A., Chambers, J.E., Uhlemann, S., Wilkinson, P.B., Meldrum, P.I., Dijkstra, T.A., Haslam, E., Kirkham, M., Wragg, J., Holyoake, S., Hughes, P.N., Hen-Jones, R. and Glendinning, S., 2014. Moisture monitoring in clay embankments using electrical resistivity tomography. *Construction and Building Materials*, pp. 1–13.
- Gunn, D.A., Ogilvy, R., Chambers, J. and Meldrum, P., 2010. ALERT-ME New technologies for embankment warning systems. *Rail Technology Magazine*, Oct/Nov., pp. 80–82.
- Hallof, P.G., 1964. A comparison of the various parameters employed in the variable frequency induced polarization method. *Geophysics*, vol. 29, pp. 425–434.
- Hassan, E., Rai, J.K. and Anekwe, U.O., 2017. Geoelectrical survey of ground water in some parts of Kebbi State Nigeria, a case study of Federal Polytechnic Bye-Pass Birnin Kebbi and Magoro Primary Health Center Fakai Local Government. *Geosciences*, vol. 7, no. 5, pp. 141–149, DOI: 10.5923/j.geo.20170705.01.
- Jakosky, J.J., 1950. *Exploration Geophysics*. Los Angeles: Trija, p. 1195.
- Joga Rao, M.V., Ghatak, S.K. and Ram Mohan, P.K., 1974. Progress report on reconnaissance gravity, magnetic and electrical resistivity surveys for coal in the Chanda-Wardha valley coal field area, Chandrapur district, Maharashtra. GSI Unpub. Rep.
- Joga Rao, M.V., Sarkar, R.K., Ghatak, T.K., Saha, D.K. and Ghosh, D.C., 1984. Report on the geophysical investigations under Deep Geology Project for delineation of Gondwana sediments below Deccan traps in parts of Yeotmal district, Wardha valley coal field area, Maharashtra. GSI Unpub. Rep.
- Jones, F., 2007. *DC Resistivity Surveys*. Washington, DC: UBC Earth and Ocean Sciences.
- Jupp, D.L.B. and Vozoff, K., 1975. Stable iterative methods of inversion of geophysical data. *Geophysical Journal RAS*, vol. 4, (Implemented by T. Harinarayana on VAX 11/750 at NGRI).
- Kaab, A. and Vollmer, M., 2000. Surface geometry, thickness changes and flow fields on permafrost streams: Automatic extraction by digital image analysis. *Permafrost Periglacial Processes*, vol. 11, no. 4, pp. 315–326.

- Kalenov, E.N., 1957. *Interpretation of Vertical Electrical Sounding Curves*. Moscow: Gostoptekhizdat, p. 472.
- Kearey, P. and Brooks, M., 1991. *An Introduction to Geophysical Exploration*. Oxford and Boston: Blackwell Scientific Publications.
- Kearey, P. and Brooks, M., 2002. *An Introduction to Geophysical Exploration*. Oxford: Blackwell Scientific Publications.
- Keller, G.V., 1966. Electrical properties of rocks and minerals. In: *Handbook of Physical Constants, Geol. Soc. Am. Mem.*, vol. 97, pp. 553–578.
- Keller, G.V. and Frischknecht, F.C., 1966. *Electrical Methods in Geophysical Prospecting*, vol. 10. Oxford and New York: Pergamon Press Inc.
- Kirsch, R., 2006. *Groundwater Geophysics: A Tool for Hydrogeology*. New York: Springer-Verlag Berlin Heidelberg, 30/2132/AO 543210, p. 489.
- Koefoed, O., 1979. *Resistivity Sounding Measurements*. New York: Elsevier Scientific Publishing co.
- Kumar, D., Ahmed, S., Krishnamurthy, N.S. and Dewandal, B., 2007. Reducing ambiguities in vertical electrical sounding interpretation: A geostatistical application. *Journal of Applied Geophysics*, vol. 62, pp. 16–32.
- Kumar, R. and Das, U.C., 1977. Transformation of dipole to Schlumberger sounding curves by means of digital linear filters. *Geophysical Prospecting*, vol. 25, pp. 780–789.
- Kunetz, G., 1966. *Principles of Direct Current Resistivity Prospecting*. Berlin: Gebruder Borntraeger, p. 103.
- LaBrecque, D.J., Ramirez, A.L., Daily, W.D., Binley, A.M. and Schima, S.A., 1996. ERT monitoring of environmental remediation processes. *Measurement Science and Technology*, vol. 7, no. 3, pp. 375–385.
- Lebourg, T., Binet, S., Tric, E., Jomard, H. and El Bedoui, S., 2005. Geophysical survey to estimate the 3D sliding surface and the 4D evolution of the water pressure on part of a deep seated landslide. *Terra Nova*, vol. 17, no. 5, pp. 399–406.
- Legault, J.M., Carriere, D. and Petrie, L., 2008. Synthetic model testing and distributed acquisition dc resistivity results over an unconformity uranium target from the Athabasca Basin, northern Saskatchewan. *The Leading Edge*, vol. 27, no. 1, pp. 46–51.
- Linderholm, P., Marescot, L., Loke, M.H. and Renaud, P., 2008. Cell culture imaging using microimpedance tomography. *IEEE Transactions on Biomedical Engineering*, vol. 55, no. 1, pp. 138–146.
- Loke, M.H., 1994. The inversion of two-dimensional resistivity data. Unpub. PhD thesis, University of Birmingham.
- Loke, M.H., 2002. Electrical imaging surveys for environmental and engineering studies: A practical guide to 2-D and 3-D surveys. *Journal of Applied Geophysics*, p. 127.
- Loke, M.H., 2004. Rapid 2-D resistivity and IP inversion using the least-square method. Manual for RES2DINV version 3.54, p. 53.
- Loke, M.H., 2015. Electrical imaging survey for environmental and engineering studies. A practical guide to 2-D and 3-D surveys, vol. 187.
- Loke, M.H. and Barker, R.D., 1996. Rapid least-squares inversion of apparent resistivity pseudosections by a quasi-Newton method. *Geophysical Prospecting*, vol. 44, pp. 131–152.
- Loke, M.H., Chambers, J.E., Rucker, D.F., Kuras, O. and Wilkinson, P.B., 2013. Recent developments in the direct-current geoelectrical imaging method. *Journal of Applied Geophysics*, vol. 95, pp. 135–156.
- Loke, M.H. and Dahlin, T., 2002. A comparison of Gauss-Newton and quasi-Newton methods in resistivity imaging inversion. *Journal of Applied Geophysics*, vol. 49, pp. 149–162, DOI: 10.1016/S0926-9851(01)00106-9.
- Loke, M.H. and Lane Jr., J.W., 2004. Inversion of data from electrical resistivity imaging surveys in water-covered areas. *Exploration Geophysics*, vol. 35, no. 4, pp. 266–271.
- Lowrie, W., 2007. *Fundamentals of Geophysics*. 2nd edn. Published in the United States of America by New York: Cambridge University Press.
- Maillet, R., 1947. The fundamental equations of electrical prospecting. *Geophysics*, vol. 3, pp. 529–556.
- Mantlik, F., Matias, M., Lourenco, J., Grangeia, C. and Tareco, H., 2009. The use of gravity methods in the internal characterization of landfills—a case study. *Journal of Geophysics and Engineering*, vol. 6, no. 4, pp. 357–364.
- Maslakowski, M., Kowalczyk, S. and Jozefiak, K., 2014. Using electrical resistivity tomography (ERT) as a tool in Geotechnical investigation of the subsurface of a highway. *Studia Quaternaria*, vol. 31, no. 2, pp. 83–89, DOI: 10.2478/squa-2014-0008.
- Matias, M. and Hermozilha, H., 2006. A 3-D constant offset GPR survey on Ilhavo Landfill. Proceedings of the 12th European Meeting of Environmental and Engineering Geophysics, Helsinki, Finland A031, 5p. DOI: <https://doi.org/10.3997/2214-4609.201402728>.

- Maurer, H. and Hauck, C., 2017. Geophysical Imaging of Alpine Rock Glaciers. *Journal of Glaciology*, Published online by Cambridge University Press, vol. 53, no. 180.
- Merritt, A.J., 2014. 4D geophysical monitoring of hydrogeology precursors to landslide activation. PhD Thesis, School of Earth and Environmental, University of Leeds, UK. p. 276.
- Muskat, M., 1933. Potentials about an electrode on the surface of the earth. *Physics*, vol. 4, pp. 129–147.
- Muskat, M. and Evinger, H.H., 1941. Current penetration in direct current prospecting. *Geophysics*, vol. 6, pp. 297–427.
- Musset, A.E. and Khan, M.A., 2000. *Looking into the Earth*. New York: Cambridge University Press.
- Naskar, D.C., 2016. Tectonic and Crustal Understanding of Lesser Himalayas along Purnea-Sevoke Transect through integrated Geophysical Studies. *Bollettino di Geofisica Teorica ed Applicata*, vol. 57, no. 4, pp. 301–312.
- Naskar, D.C., 2017. Geophysical surveys over Bakreswar geothermal region in Birbhum district, West Bengal. *Journal of Indian Geophysical Union*, vol. 21, no. 5, pp. 343–352.
- Naskar, D.C., Choudhury, K., Roy, K.K., Singh, O.P. and Majumdar, R.K., 2002. Geophysical approach for delineation of shallow crustal structure along Bargaon-Sanwer transect, Madhya Pradesh. *Journal of Geological Society of India*, vol. 60, pp. 173–181.
- Naskar, D.C., Das, L.K. and Lahiri, A.K., 2008. Integrated geophysical surveys for delineation of coal bearing Gondwana rocks in Ulia–Hargawan area, Surguja district, Chhatisgarh. *Journal of Geological Society of India*, vol. 71, pp. 89–97.
- Naskar, D.C. and Das, L.K., 2009. Resistivity survey across the Narmada-Son lineament and along the Sanwer-Borgaon traverse, Madhya Pradesh, central India. *Bollettino di Geofisica Teorica ed Applicata*, vol. 50, no. 4, pp. 305–318.
- Naskar, D.C., Majumdar, R.K., Choudhury, K., Roy, K.K. and Das, L.K., 2006. Results of deep geoelectric sounding over Deccan trap of Madhya Pradesh. *Journal of Geophysics*, vol. XXVII, no. 4, pp. 81–90.
- Naskar, D.C. and Saha, D.K., 2015. Geophysical investigations for delineation of Gondwana sediments below Deccan trap beyond the Western limit of Wardha Valley Coalfields, Yeotmal and Wardha Districts, Maharashtra-a comprehensive analysis of case studies. *Journal of Indian Geophysical Union*, vol. 19, no. 4, pp. 433–446.
- Naskar, D.C. and Rai, M.K., 2018. Delineation of geometry and spatial distribution of basic bodies using magnetic and resistivity methods in Sohagpur coal field area, Madhya Pradesh. *International Journal of Research-Granthaalayah*, vol. 6, no. 8, pp. 146–158, ISSN 2350-0530(O), ISSN 2394-3629(P), DOI: 10.5281/zenodo.1403846.
- Naskar, D.C., Singh, O.P., Das, L.K. and Syiem, S.M., 2011. Integrated geophysical survey for delineation of active fault in the foothill of Bhutan Himalaya, northeast India. *Indian Journal of Geosciences*, vol. 65, no. 3, pp. 211–222.
- National Research Council, 1996. *Rock Fractures and Fluid Flow: Contemporary Understanding and Applications*. Washington, DC: National Acad. Press.
- Nimmer, R.E. and Osiensky, J.L., 2002. Using mise-a-la-masse to delineate the migration of a conductive tracer in partially saturated basalt. *Environmental Geosciences*, vol. 9, pp. 81–87.
- Olayinka, A.I. and Mbachi, C.N.C., 1992. A technique for the interpretation of electrical sounding from crystalline basement Areas of Nigeria. *Journal of Mining Geology*, vol. 27, pp. 63–69.
- Olorunniwo, M.A. and Olorunfemi, M.O., 1987. Geophysical investigation for groundwater in Precambrian terrain, a case history from Ikare. Southwestern Nigeria. *Journal Afr. Earth Sciences*, vol. 6, pp. 787–796.
- Orellana, E., 1960. Algunas cuestiona de prospeccion geoelectrica. *Revista de Geofisica*, vol. 19, pp. 13–28.
- Orellana, E., 1961. Criterios erroneos en la interpretation de sondeos electricos. *Revista de Geofisica*, vol. 20, pp. 207–227.
- Orellana, E. and Mooney, H.M., 1966. Master tables and curves for vertical electrical sounding over layered structures. *Madrid Interciecia*, p. 166, 66 tables.
- Parasnis, D.S., 1962. *Principles of Applied Geophysics*. London: Methuen and Co. Ltd., p. 176.
- Petersen, T. and al Hagry, S.A., 2009. Mapping root zones of small plants using surface and borehole resistivity tomography. *Leading Edge*, vol. 28, no. 10, pp. 1220–1224.
- Rahmani, T.R., Sari, D.P., Akam, A., Amir, H. and Putra, A., 2020. Using the Schlumberger configuration resistivity geoelectric method to analyze the characteristics of slip surface at Solok. The 2nd International Conference on Research and Learning of Physics IOP Conf. Series: Journal of Physics: Conf. Series 1481 (2020) 012030 IOP Publishing.

- Reynolds, J.M., 1997. *An Introduction to Applied and Environmental Geophysics*. Chichester: John Wiley and Sons.
- Reynolds, J.M., 1998. *An Introduction to Applied and Environmental Geophysics*. New York: Wiley.
- Reynolds, J.M., 2011. Electrical resistivity sounding and tomography. Technical Summary Technical Summary. *Reynolds International*, vol. 2, p. 2.
- Robinson, E.S. and Coruh, C. 1988. *Basic Exploration Geophysics*. New York: John Wiley and Sons.
- Roy, K.K., Das, L.K., Das, H., Mukherjee, K.K., Sen, M.K., Routh, P.S., Chunduru, R., Saha, D.K., Rai, M.K. and Naskar, D.C., 1997. Direct current resistivity traversing across Singhbhum shear zone near Ghatsila-Mosabani, eastern India. *Journal of Geophysics*, vol. XVIII, no. 4, pp. 215–221.
- Roy, K.K., Das, L.K., Das, H.K., Mukherjee, K.K., Sen, M.K., Routh, P.S., Saha, D.K., Rai, M.K., Naskar, D.C. and Chunduru, R., 1998. Direct current resistivity traversing across Singhbhum shear zone near Ghatsila-Mosabani, Eastern India. In Roy, K.K., et al. (eds), *Deep Electromagnetic Exploration*, copyright © 1998. New Delhi: Narosa Publishing House.
- Schlumberger, C., 1920. *Etude sur la prospection électrique du sous-sol*. Paris: Gauthier—Villars.
- Seladji, S., Cosenza, P., Tabbagh, A., Ranger, J. and Richard, G., 2010. The effect of compaction on soil electrical resistivity: A laboratory investigation. *European Journal of Soil Science*, vol. 61, pp. 1043–1055.
- Silvester, P.P. and Ferrari, R.L., 1990. *Finite Elements for Electrical Engineers*. 2nd edn. Cambridge University Press.
- Sorensen, K.I., 1996. Pulled Array Continuous Profiling. *First Break*, vol. 14, pp. 85–90.
- Souplos, P., Papadopoulos, I., Kouli, M., Georgaky, I., Vallianatos, F. and Kokkinou, E., 2007. Investigation of waste disposal areas using electrical methods: A case study from Chania, Crete, Greece. *Environmental Geology*, vol. 51, pp. 1249–1261.
- Stampolidis, A., Tsourlos, P., Souplos Mimides, P., Tsokas, G., Vargemezis, G. and Vafidis, A., 2005. Integrated geophysical investigation around the brackish spring of Rina, Kalimnos Isl., SW Greece. *J. Balk. Geophysical Society*, vol. 8, pp. 63–73.
- Storz, H., Storz, W. and Jacobs, F., 2000. Electrical resistivity tomography to investigate geological structures of the Earth's upper crust. *Geophysical Prospecting*, vol. 48, no. 3, pp. 455–471.
- Stummer, P., Maurer, H.R., Horstmeyer, H. and Green, A.G., 2002. Optimization of DC resistivity data acquisition: Real-time experiment design and a new multi-electrode system. *IEEE Trans. Geosci. Remote Sens.*, vol. 40, no. 12, pp. 2727–2736.
- Sumner, J., 1976. *Principles of Induced Polarization for Geophysical Exploration*. New York: Elsevier, p. 277.
- Tabbagh, A., Dabas, M., Hesse, A. and Panissod, C., 2000. Soil resistivity: A non-invasive tool to map soil structure horization. *Geoderma*, vol. 97, pp. 393–404.
- Telford, W.M., Geldart, L.P. and Sheriff, R.E., 1990. *Applied Geophysics*. 2nd edn. UK: Cambridge University Press, p. 770.
- Telford, W.M., Geldart, L.P., Sheriff, R.E. and Keys, D.A., 1976. *Applied Geophysics*. Cambridge: Cambridge University Press.
- Van Nostrand, R.G. and Cook, K.L., 1966. Interpretation of Resistivity Data, Prof. Paper No. 499, United States Geological Survey.
- Verheijen, P.J.T. and Ajakaiye, D.E., 1979. Geophysical anomalies over a pipe suspectedly kimberlite in the Precambrian metamorphic schist belt of Northern Nigeria. *Geoexploration*, vol. 17, pp. 293–303. © Elsevier Scientific Publishing Company, Amsterdam—Printed in The Netherlands.
- Ward, S.H., 1990. Resistivity and induced polarization methods. *Journal of Geotechnical and Environmental Geophysics*, vol. 1, Ward, S.H., Ed: Society of Exploration Geophysicists.
- Weaver, W., 1929. Certain face potential method. *Trans.*, vol. 81, pp. 68–86.
- Wenner, F., 1915. A method for measuring earth resistivity. *Journal of the Franklin Institute (Elsevier)*, vol. 180, no. 3, pp. 373–375.
- Wilkinson, P.B., Loke, M.H., Meldrum, P.I., Chambers, J.E., Kuras, O., Gunn, D.A. and Ogilvy, R.D., 2012. Practical aspects of applied optimized survey design for electrical resistivity tomography. *Geophysical Journal International*, vol. 189, no. 1, pp. 428–440.
- Wynn, J. and Sherwood, S., 1984. The self-potential (SP) method—an inexpensive reconnaissance and archaeological mapping tool. *Journal of Field Archaeology*, vol. 11, pp. 195–204.
- Xu, B. and Noel, M., 1993. On the completeness of data sets with multi-electrode systems for electrical resistivity survey. *Geophysical Prospecting*, vol. 41, pp. 791–801.
- Yadav, G.S. and Singh, S.K., 2007. Integrated resistivity surveys for delineation of fractures for groundwater exploration in hard rock areas. *Journal of Applied Geophysics*, vol. 62, pp. 301–312.

- Zhou, B. and Dahlin, T., 2003. Properties and effects of measurement errors on 2-D resistivity imagine survey. *Near Surface Geophysics*, vol. 1, no. 3, pp. 105–117.
- Zohdy, A.A.R., 1964. Earth resistivity and seismic refraction investigations in Santa Clara County, California. Unpub. Ph.D. thesis, Stanford Univ. p. 132.
- Zohdy, A.A.R., 1965a. The auxiliary point method of electrical sounding interpretation, and its relationship to the Dar Zarrouk parameters. *Geophysics*, vol. 30, pp. 644–660.
- Zohdy, A.A.R., 1968b. The effect of current leakage and electrode spacing errors on resistivity measurements, in Geological Survey Research 1968. U.S. Geol. Survey Prof. Paper 600-D, pp. D268-D264.
- Zohdy, A.A.R., 1969a. The use of Schlumberger and equatorial soundings in ground-water investigations near El Paso, Texas. *Geophysics*, vol. 34, pp. 713–728.
- Zohdy, A.A.R., 1969b. A new method for differential resistivity sounding. *Geophysics*, vol. 34, pp. 924–943.
- Zohdy, A.A.R., 1970. Variable Azimuth Schlumberger resistivity sounding and profiling near a vertical contact: New techniques in direct current resistivity exploration. *United State Geological Survey Bulletin*, no. 1313, pp. A1–A22.
- Zohdy, A.A.R., Eaton, G.P. and Mabey, D.R., 1964. Application of surface geophysics to groundwater investigations. Chapter D1, USGS. Techniques of Water-Resources Investigations of the United States Geological Survey. Book 2, collection of environmental data, p. 66.
- Zohdy, A.A.R., Eaton, G.P. and Mabey, D.R., 1984. Application of Surface Geophysics to Ground Water Investigations. United States Department of the Interior, Book 2, Chapter D1, pp. 1–66.
- Zohdy, A.A.R. and Jackson, D.B., 1969. Application of deep electrical soundings for ground water exploration in Hawaii. *Geophysics*, vol. 34, pp. 584–600.
- Zohdy, A.A.R., Jackson, D.B., Mattick, R.E. and Peterson, D.L., 1969. Geophysical Surveys for Ground Water at White Sands Missile Range, New Mexico. U.S. Geol. Survey open file Rep. p. 31.

6 Natural Potentials and Currents

Electrical investigations of natural electrical properties are based on the measurement of the voltage between a pair of electrodes implanted in the ground. Natural differences in potential occur in relation to subsurface bodies that create their own electric fields. The bodies act like simple voltaic cells; their potential arises from electrochemical action. Natural currents (called telluric currents) flow in the crust and mantle of the Earth. They are induced electromagnetically by electric currents in the ionosphere. In studying natural potentials and currents, scientists have no control over the source of the signal. This restricts the interpretation, which is mostly only qualitative. The natural methods are not as useful as controlled induction methods, such as resistivity and electromagnetic techniques, but they are inexpensive and fast.

6.1 SELF-POTENTIAL (SP) METHOD: INTRODUCTION

A potential that originates spontaneously in the ground is called a *self-potential* (or *spontaneous potential*). Some self-potentials are due to man-made disturbances of the environment, such as buried electrical cables, drainage pipes, or waste disposal sites. They are important in the study of environmental problems. Other self-potentials are natural effects due to mechanical or electrochemical action. In every case, the groundwater plays a key role by acting as an electrolyte.

Some self-potentials have a mechanical origin. When an electrolyte is forced to flow through a narrow pipe, a potential difference (voltage) may arise between the ends of the pipe. Its amplitude depends on the electrical resistivity and viscosity of the electrolyte and on the pressure difference that causes the flow. The voltage is due to differences in the electrokinetic or streaming potential, which in turn is influenced by the interaction between the liquid and the surface of the solid (an effect called the zeta potential). The voltage can be positive or negative and may amount to some hundreds of millivolts. The electrokinetic potential (E_k) generated between the ends of the capillary passage is given by

$$E_k = \frac{\epsilon \rho C_E \Delta P}{4\pi\eta}$$

where ϵ is the dielectric permittivity of pore fluid, ρ is the electrical resistivity of pore fluid, C_E is the electrofiltration coupling coefficient, ΔP is the pressure difference, and η is the dynamic viscosity of pore fluid. This type of self-potential can be observed in conjunction with seepage of water from dams or the flow of groundwater through different lithological units, over terrains with large elevation changes, and in geothermal areas.

Some self-potentials have a thermoelectric origin. The potential gradient will appear across a rock sample if a temperature gradient is maintained across the rock sample (thermoelectric [TE] potential). The *thermoelectric coupling coefficient* (TEC) is defined as the ratio of the voltage to the temperature difference:

$$\text{TEC} = \frac{\Delta V}{\Delta T}$$

TEC values of rocks vary from -0.09 to $+1.36$ mV/°C and average ~ 0.27 mV/°C. SP generated from TE potentials are of smaller amplitudes than usually seen in geothermal areas. More concentrated areas of high temperature at shallow depth, such as thermal fluids in a fault zone, could give rise to anomalies of greater amplitude. Boundaries of SP anomalies measured in several geothermal

areas appear to correlate with zones of known anomalous high heat flow—a portion of anomalies is generated by the TE mechanism.

Most self-potentials have an electrochemical origin. For example, if the ionic concentration in an electrolyte varies with location, the ions tend to diffuse through the electrolyte so as to equalize the concentration. The diffusion is driven by an electric diffusion potential, which depends on the temperature as well as the difference in ionic concentration. When a metallic electrode is inserted in the ground, the metal reacts electrochemically with the electrolyte (i.e. groundwater), causing a contact potential. If two identical electrodes are inserted in the ground, variations in the concentration of the electrolyte cause different electrochemical reactions at each electrode. A potential difference arises, called the Nernst potential. The combined diffusion and Nernst potentials are called the *electrochemical self-potential* (Wikipedia, 2020). It is temperature-sensitive and may be either positive or negative, amounting to, at most, a few tens of millivolts. Electrochemical potential (E_C) is the sum of liquid junction or diffusion potential (E_J) and Nernst or membrane potential (E_M):

$$E_C = E_M + E_J = K \log_{10} \frac{a_w}{a_{mf}}$$

where K is constant and varies in direct proportion with temperature, especially in NaCl solutions ($K = 61 + 0.133T$ in °F, and $K = 65 + 0.24T$ in °C), a_w is the formation water ionic activity, and a_{mf} is the mud filtrate ionic activity. Since spontaneous potential (SP) is a measure of electrochemical potential and the ionic activity of a solution is inversely proportional to its resistivity, the previous equation can be simplified as

$$SP = E_C = K \log_{10} \frac{\rho_{mfe}}{\rho_{we}}$$

where ρ_{mfe} is the equivalent mud filtrate resistivity and ρ_{we} is the equivalent formation water resistivity. The ideal spontaneous potential across a clean bed is known as static SP (SSP) and is defined as follows:

$$SSP = -K \log_{10} \frac{\rho_{mfe}}{\rho_{we}}$$

The self-potentials that originate by the previous mechanisms are attracting increased attention in environmental and engineering situations. However, in the exploration for subsurface regions of mineralization, they are often smaller than the potentials associated with ore bodies and are classified accordingly as ‘background potentials’. The self-potential associated with an ore body is called its ‘mineralization potential’. Self-potential (SP) anomalies across ore bodies are invariably negative, amounting usually to a few hundred millivolts. They are most commonly associated with sulphide ores, such as pyrite, pyrrhotite, and chalcopyrite, but also with graphite and some metallic oxides.

The origin of the mineralization type of self-potential is still obscure, despite decades of applied investigations. At one time, it was thought that the effect arose from galvanic action. This occurs when dissimilar metal electrodes are placed in an electrolyte. Unequal contact potentials are formed between the metals and the electrolyte, giving rise to a potential difference between the electrodes. According to this model, an ore body behaves like a simple voltaic cell, with groundwater acting as the electrolyte. It was believed that oxidation of the part of the ore body above the water table produced a potential difference between the upper and lower parts, causing a spontaneous electric polarization of the body. Oxidation involves the addition of electrons so the top of the ore body becomes negatively charged, explaining the observed negative anomalies. Unfortunately, this simple model does not explain many of the observed features of self-potential anomalies and has proved to be untenable.

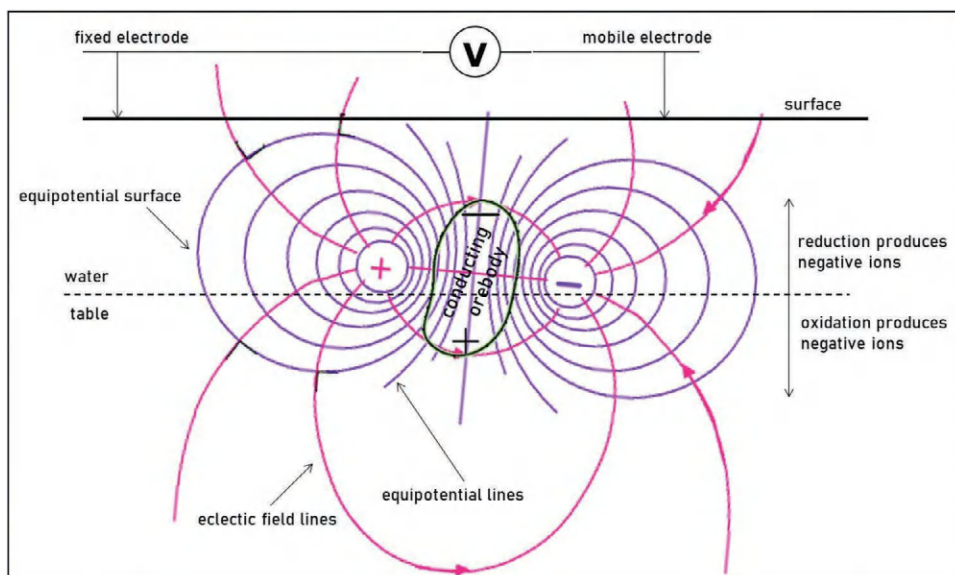


FIGURE 6.1 A schematic model of the origin of the SP anomaly of an ore body. The mechanism depends on differences in oxidation potential above and below the water table.

Another mechanism for self-potential depends on variations in oxidation (redox) potential with depth (Figure 6.1). The ground above the water table is more accessible to oxygen than the submersed part, so moisture above the water table contains more oxidized ions than that below it. An electro-chemical reaction takes place at the surface between the ore body and the host rock above the water table. It results in the reduction of the oxidized ions in the adjacent solution. An excess of negative ions appears above the water table. A simultaneous reaction between the submersed part of the ore body and the groundwater causes oxidation of the reduced ions present in the groundwater. This produces excess positive ions in the solution and liberates electrons at the surface of the ore body, which acts as a conductor connecting the two half-cells (zone of aeration and zone of saturation). Electrons flow from the deep part to the shallow part of the ore body. Outside the ore body, positive ions move from bottom to top along the electric field lines. The equipotential surfaces are normal to the field lines. The self-potential is measured where they intersect the ground surface (Figure 6.1) (Sato and Mooney, 1960).

The redox model is inadequate for the same reason as the galvanic model: it fails to account for many of the observed features of self-potential anomalies. In particular, the association of self-potential models with the water table has been cast in doubt. Moreover, sulphide ore bodies appear to persist for geological lengths of time, so a mechanism involving the permanent flow of charge appears unlikely. *Self-potential* is a feature of a stable system that is perturbed by making an electrical connection between the host rock and the sulphide conductor through the inserted electrodes and their connecting wire. The observed potential difference appears to be due to the difference in oxidation potential between the locations of the measurement electrodes, one inside and the other outside the zone of mineralization.

The following table (Table 6.1) shows the SP anomalies and their geological sources.

6.2 SURVEY EQUIPMENT

Survey equipment comprises of sensitive high-impedance digital DC voltmeter, a pair of nonpolarizing electrodes (Cu-CuSO_4), and a sufficiently long cable (required tensile strength to withstand pulls, drags, and light enough to carry along the traverse).

TABLE 6.1**Types of SP Anomalies and Their Geological Sources**

Source	Type of Anomaly
Mineral Potentials	
Sulphide ore bodies (pyrite, chalcopryrite, pyrrhotite, sphalerite, galena)	Negative $\approx$ hundreds of mV
Graphite ore bodies	
Magnetite + other electronically conducting minerals	
Coal	
Manganese	
Quartz veins	Positive $\approx$ tens of mV
Pegmatites	
Background Potentials	
Fluid streaming, geochemical reactions, etc.	Positive/negative, ≤ 100 mV
Bioelectric (plants, trees)	Negative, ≤ 300 mV or so
Groundwater movement	Positive/negative), up to hundreds of mV
Topography	Negative, up to 2 V

6.3 MEASUREMENT OF SP/FIELD PROCEDURE

As far as possible, traverses are run perpendicular to the strike. Normally, the station interval is about 10 m. Small pits are prepared before the survey near each station and are watered for good electrical contact. Preferably, watering should be done the previous evening to avoid small potential fluctuations due to electrofiltration. Measurement of SP is simple and inexpensive.

Porous pots are filled with a uniform batch of solution interconnected electrically and then placed in a tub of the same solution and left overnight. This will minimize the potential difference between the two pots, which should desirably be zero. For this reason, it is preferable to make the solution with distilled water. Usually, several such pots are prepared, and the identical pairs are chosen for field observations. Two nonpolarizable porous pot electrodes are connected to precision voltmeters capable of measuring to at least 1 mV. Two electrodes are placed in the ground a couple of metres apart, with one electric wire running from each electrode to the instrument. When the user takes a reading, the instrument reads the voltage difference between the two electrodes and stores the value in its memory. Each electrode is made up of a copper electrode dipped in a saturated solution of copper sulphate, which can percolate through the porous base to the pot (Bogdan et al., 2012). An alternate zinc electrode in saturated zinc sulphate solution, or silver in silver chloride, can be used. The maximum depth of sensitivity of the SP method ≈ 60 –100 m, depending on the ore body and the nature of the overburden. Two field techniques or two electrode configurations are normally used in the SP survey.

6.3.1 POTENTIAL GRADIENT METHOD (DIPOLE/LEAP-FROG/GRADIENT CONFIGURATION)

Fix separation of two electrodes (5 or 10 m). Measure the potential difference between two electrodes = potential gradient (mV/V). Two porous pots are leap-frogged along the traverse with care of correct polarity of potential recorded (Figure 6.2a). Observation points = midpoint between two electrodes. The benefit of the gradient method is that even if you measure 1,000 m, you only need a 10 m wire between the two electrodes. The issue with the gradient method is that you get an error with each reading. So when you sum all the gradient measurements up, it adds to the error. This tends to hide any small anomalies.

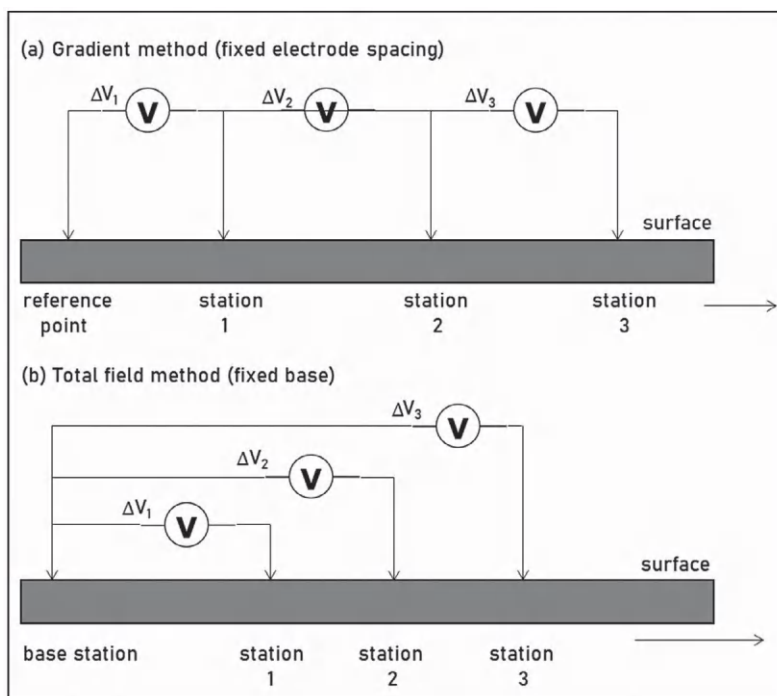


FIGURE 6.2 The field techniques of measuring self-potential by (a) the gradient method and (b) the total field method. The total potential V at a station in the gradient method is found by summing the previous potential differences ΔV ; in the total field method, V is measured directly.

6.3.2 POTENTIAL AMPLITUDE, OR TOTAL FIELD METHOD (FIXED-BASE) CONFIGURATION

Keep one electrode fixed at a base station (Figure 6.2b) without moving it. This electrode is called the base station and acts as the reference point for all other measurements (Lowrie, 2007). To follow the fixed-base SP method, you simply move the other electrode a certain distance forward (say, 10 m) along a line and take a reading. You then move it another 10 m forward, so it's 20 m from the base station, and repeat this process until the end of the line. The result of the survey is plotted against the distance from the base station. Measure the potential difference (mV) between the base and the second electrodes moving along traverse. There is a lower level of cumulative errors and confusing polarity. Disadvantages include the transport of long wire. Strong negative anomalies indicate mineralization.

In the second method of traversing, the use of very long cables should be avoided, even if the traverse to be run may be considerably long. This is not only because large cables may introduce its own resistance and thereby introduce error but also because the telluric potentials may interfere with the SP values. Also, there are logistical difficulties with large cables, especially in wooded and cultured areas. Not more than a kilometre of cable is used at a time, and the base station is shifted to a new location. This new base station is accurately tied to the earlier base station by noting the SP between the base stations, and further observations on traverses are recorded with respect to the new 'base'.

The porous pots do normally develop some potential difference between them during the course of the surveys, owing to temperature changes or other causes. These changes are accounted for by a determination of the change in potential difference at the beginning and end of the surveys and by linearly distributing errors at various stations to obtain corrected values. Alternatively, at the end of a set of observations, both base and moving pots are interchanged, and the measuring stations are reoccupied. The average value of a station is taken for data plotting. In conducting any SP surveys, care should always be taken to maintain the same polarity of the electrode. As a rule, the moving electrode

on the traverse is recorded positive and kept at base as negative. Balancing the galvanometer with the commutator at normal means that the SP value is positive, and at reverse means negative.

The recognition of different spontaneous potential sources is important to eliminate noise, the low background voltages. Some engineering and environmental occurrences may be mapped by contouring surficial voltages between the base/reference electrode(s) and the mobile electrodes. Flow of gasses and fluids in pipes, leakage of a reservoir within the foundation or abutment of a dam, movement of ionic fluids to or within the groundwater, flow of geothermal fluids, and movement of water into or through a karst system can be the origin of streaming potentials. These potentials may exceed the background voltage variation of a site.

6.4 INTERPRETATION OF SP DATA

SP anomalies are often interpreted qualitatively by profile shape, amplitude, polarity (positive or negative), and contour pattern. The series of voltage readings helps the user see if the ground is homogeneous (and thus has the same value) or if there is an SP anomaly detected. The user is able to tell if an anomaly—caused by an underground sulphide ore body—is detected, because the voltmeter will register a major dip in voltage; the anomalies from an ore body are always negative over a span of the ground that has been measured. If you just measure on regular ground, the anomalies are typically in the order of a few millivolts. When you are measuring over an ore body, you can get negative voltages of several hundred millivolts and down to a volt for a large anomaly. The cause of the negative effect of sulphide ore bodies is complex and not fully understood. However, the negative anomaly reading is likely caused by oxidation at the top of the ore body, which is created through oxygen-rich rainwater percolating downward. This causes a difference in polarity at the top and bottom of the ore body and creates what is essentially a huge but weak battery on the Earth. For those interested in mineral exploration, it is worth noting that detecting a large negative anomaly is not necessarily indicative of an ore body—it could also be graphite. One is clearly more valuable than the other, though graphite can still be a great find. The top of the ore body is assumed to lie directly beneath the position of minimum potential (Figure 6.3).

For quantitative interpretation, it is possible to calculate the potential distributions around polarized bodies of simple shape, such as sphere, ellipsoid, and dipole, by making some simplifications and assumptions concerning the potential on the surface of the sources. Some examples of SP anomalies are given for easy understanding (Stern, 1945; Petrovski, 1928).

6.5 APPLICATIONS OF SELF-POTENTIAL (SP) METHOD WITH SOME CASE HISTORIES

The SP geophysical method has been used for a variety of purposes, from exploration for ores (Corry, 1985) and hydrocarbons (Atekwana and Atekwana, 2010; Frasheri, 2002), to archaeological mapping (Wynn and Sherwood, 1984), to studies of groundwater flow (Reynolds, 1997). Recently, it has been recommended for tracing contaminant spills in environmental remediation (Nyquist and Corry, 2002; Park et al., 2004; Howells and Fox, 1998; Corwin, 1990). The appeals of the method are the simplicity of the survey procedure, the low cost, and the straightforward nature of the signal: an anomaly occurs directly above its source, and the width of an anomaly is related to its extent and its burial depth. The major drawback of the method is that there are many possible sources for SP anomalies, and it is sometimes difficult to know which applies. For this reason, SP surveys typically are combined with other methods, such as resistivity (Panthulu et al., 2001), seismic surveys (Davenport et al., 1978), or VLF EM, or they are included in surveys utilizing multiple techniques (e.g. Howells and Fox, 1998).

The most useful SP component is the *electrochemical potential*, since it can cause a significant deflection opposite permeable beds. The magnitude of the deflection depends mainly on the salinity contrast between the borehole and the formation fluid and the clay content of the permeable bed. The SP log is therefore useful in detecting permeable beds and estimating formation water salinity and formation clay content. Due to the nature of the electric current, SP can only be recorded in conductive mud.

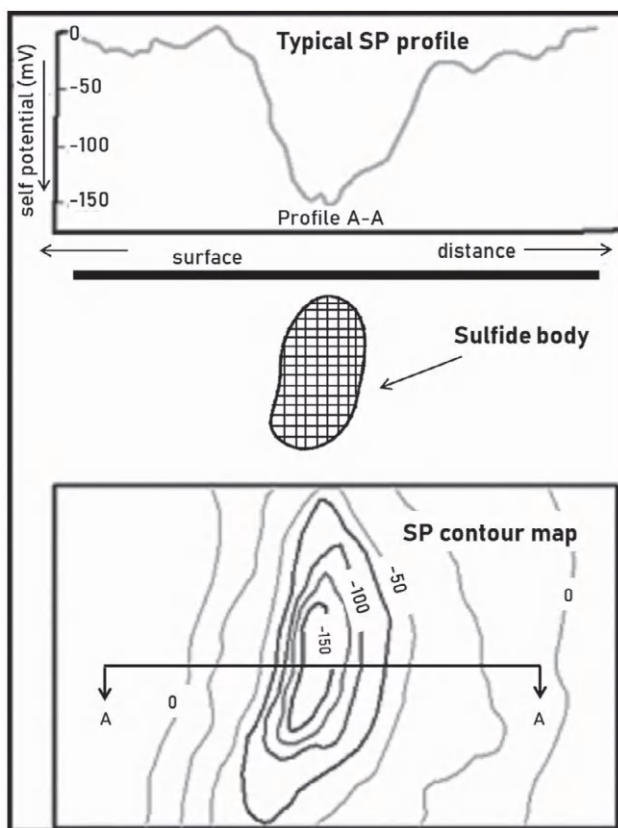


FIGURE 6.3 Hypothetical contour lines of a negative self-potential anomaly over an ore body. Negative maximum lies directly over the sulphide body.

Source: Parasnis (1966).

The SP method is primarily used in mineral (base metal) exploration. The results of SP reveal that the geochemical copper and gold anomalies coincide with the SP anomalies. A number of drill hole sites are recommended from the interpretation of the SP response to test the mineralization in the subsurface (after Naskar et al., 2018; Naskar and Kulkarni, 2019). Complications arise when two or more geological features give rise to superimposed SP anomalies (Nayak, 1981; Meiser, 1962). The SP method has also been extended to groundwater (Nordiana et al., 2017; Teh Saufia et al., 2013) and geothermal investigations (Corwin and Hoover, 1979; Corwin et al., 1979; Sill and Bodell, 1977) as shown in Figure 6.4. Steep-sided landfills containing significant volumes of highly conductive leachate that may leak through the margins are known to generate significant SP anomalies (Coleman, 1991). The SP method has been used to detect leaks in Earth dam embankments for decades (Bogoslovsky and Ogilvy, 1970a, 1970b). The concept of the generation of SP anomalies by leaks is shown schematically in Bogoslovsky and Ogilvy (1970a, 1970b). Negative charges are associated with locations where leaks enter a dam or above seepage paths, where the flow is generally horizontal or descending. In contrast, positive anomalies may occur where the flow is generally ascending and where surface seepage takes place. Consequently, the presence of either or both a negative and/or positive anomaly may be physically significant. Butler et al. (1990) have described an example of an SP survey to examine possible leaks at the Mill Creek Dam and Reservoir, Washington, USA (Butler et al., 1990).

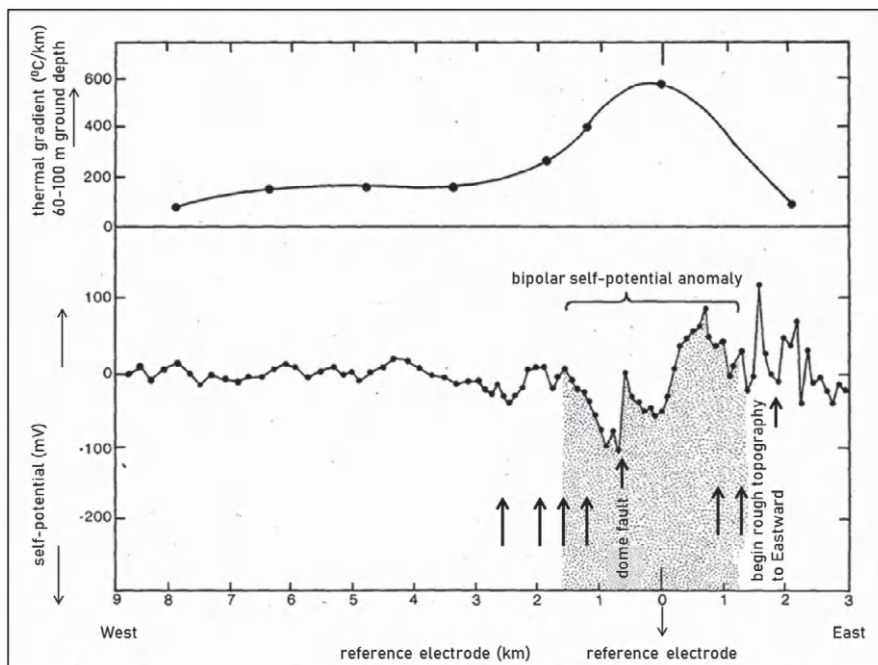


FIGURE 6.4 Thermal gradient and self-potential profile across the dome fault, Roosevelt Hot Springs, Utah, USA. Arrows denote points at which faults mapped by a geological survey cross the self-potential survey line.

Source: Thermal gradient data from Sill and Bodell (1977).

6.6 SELF-POTENTIAL AS A TOOL TO ASSESS GROUNDWATER FLOW IN HYDROTHERMAL SYSTEMS: A REVIEW OF VOLCANOLOGY MAPPING

In hydrothermal systems associated with volcanic areas, the flow of groundwater generates an electrical current of an electrokinetic nature, the so-called streaming current. In the context of the exploration of such hydrothermal reservoirs, geophysical methods play a critical role. The self-potential method is a simple, nonintrusive, and passive geophysical method that is considered to be one of the oldest methods in geophysics (Fox, 1830; Ahmad, 1969; Bogoslovsky and Ogilvy, 1973; Sill, 1983). One specific advantage of the self-potential method is that it can be used to assess non-intrusively the flow of the groundwater (Keller, 1981; Spies and Frischknecht, 1992; Vozoff, 1992; Huenges, 2010; Jolie et al., 2021). It has also a long history in the study of hydrothermal systems (Poldini, 1938, 1939; Zohdy et al., 1973; Anderson and Johnson, 1976). There are no other geophysical methods that can be used as such a nonintrusive groundwater flow sensor.

In the realm of hydrogeophysics, the self-potential method can be used to detect groundwater flow (Poldini, 1938, 1939; Revil et al., 2005a, 2005b; Suski et al., 2006), with applications to dams and embankments (Bogoslovsky and Ogilvy, 1973; Gex, 1980; Boleve et al., 2011) and pumping tests in wells (Semenov, 1980; Malama et al., 2009a, 2009b; Soueid et al., 2016).

6.6.1 VOLCANOES IN JAPAN: MISTI VOLCANO IN PERU, VOLCANO IN MT FUJI, AND VOLCANO IN MIYAKEJIMA

For volcanoes, examples of the self-potential profiles and maps are presented in Revil et al. (2023), which has been modified from Finizola et al. (2004) for the Misti volcano in Peru and Mt Fuji and Miyakejima volcanoes in Japan (Aizawa, 2000; Aizawa, 2004; Aizawa et al., 2005). For these volcanoes and others, large-scale anomalies of hundreds of millivolts can be observed over several kilometres (Ishido et al., 1997; Aubert et al., 1990; Aizawa, 2000; Zlotnicki and Nishida, 2003; Finizola

et al., 2004; Barde-Cabusson et al., 2009b; Barde-Cabusson et al., 2012; Villasante-Marcos et al., 2014; Gonzales et al., 2014; Gresse et al., 2021). Self-potential anomalies can be related to the flow of the groundwater in shallow unconfined aquifers (Bogoslovsky and Ogilvy, 1973; Fournier, 1989; Suski et al., 2004, 2006) in the vadose zone (Aubert and Dana, 1994), along permeable paths in faults (Aubert and Dana, 1994; Revil and Pezard, 1998; Finizola et al., 2002, 2003; Hase et al., 2005; Bennati et al., 2011), or through free convection in high-permeability reservoirs (Revil et al., 1999a, 1999b). Transient self-potential signals can also be recorded with a network of nonpolarizing electrodes in the form of time change of the electrical field (Moore and Glaser, 2007; Barde-Cabusson et al., 2009a; Zlotnicki and Nishida, 2003; Matsushima et al., 2017). These fluctuations can be related to changes in groundwater flow associated with thermal or thermohydromechanical disturbances, for example, through thermal pressurization (Haas et al., 2013; Kuwano et al., 2015; Chaput et al., 2019). The self-potential method can therefore complement seismological data to decipher and monitor the plumbing system of hydrothermal fields (Mahardika et al., 2012).

6.6.2 LARGE-SCALE STUDY ON PERUVIAN AND GUATEMALAN VOLCANOES

The large-scale surveys performed on active Peruvian volcanoes, such as the Misti (Finizola et al., 2004) and Ubinas volcanoes (Gonzales et al., 2014), revealed large amplitude self-potential anomalies with a classical ‘W’ shape underlying a huge hydrothermal system several kilometres in diameter. Such a signal means the presence at a depth of a few kilometres of a magmatic body (Revil et al., 1999b, for a description of the numerical modelling case study). The density of active volcanoes in Guatemala implies particular strategy to study with self-potential, volcanic complexes such as the Santa Maria–Cerro Quemado–Zunil and the Xela caldera (Revil et al., 2023; modified from Bennati et al., 2011). A unique self-potential profile of about 17 km long, with a step of 20 m, has been performed, beginning at the top of the Santa Maria volcano, crossing the Xela caldera boundary, then following with the Cerro Quemado dome complex, and crossing again the Xela caldera boundary in a direction towards the Zunil volcano. The hydrothermal system of the Santa Maria volcano is revealed by a positive self-potential/elevation gradient (anomalies M1–M2 in Revil et al., 2023; modified from Bennati et al., 2011), and the western flank of the Xela caldera boundary also facilitates the uprising of hydrothermal fluids, as attested by the positive self-potential/elevation gradient (M4 in Revil et al., 2023; modified from Bennati et al., 2011). A similar hydrothermal system appears close to the 1818 lava dome (anomaly Q4 in Revil et al., 2023; modified from Bennati et al., 2011). Around the Cerro Quemado dome complex, the presence of negative self-potential anomalies indicates the possibility of preferential groundwater infiltration along the faults bordering the Cerro Quemado dome complex (Revil et al., 2023; modified from Bennati et al., 2011). This survey highlights the notion of partitioned hydrothermal systems on a volcanic complex, a notion attested by the isotopic gas geochemistry (Revil et al., 2023; modified from Bennati et al., 2011).

6.6.3 SMALL-SCALE STUDY IN THE SUMMIT PART OF STROMBOLI VOLCANO

A high-resolution survey (1 m spacing along profiles for self-potential and 2.5 m spacing for temperature recorded at a depth of 30 cm) was performed in 1999 in the summit area of the Stromboli volcano (Finizola et al., 2003). Revil et al. (2004) performed an electrical resistivity tomography with 5 m spacing between the electrodes. An eruptive crisis occurred in 2002–2003, which opened fractures in the summit area (Revil et al., 2023; modified from Finizola et al., 2009). It is interesting to note the strong correlation between the positive anomalies of self-potential, the temperature anomalies at a depth of 30 cm, and the conductive bodies shown on the resistivity profiles. The superimposition between the 2002–2003 opened fractures and the maximum of the self-potential anomaly (F1 in Figure 16 in Revil et al., 2023) highlights that self-potential signals can be useful to anticipate weakness planes in the structure on a volcanic edifice, which can be reactivated during an unrest period. Similar observations were made by Chaput et al. (2019) before the collapse of the Dolomieu caldera on the Piton de la Fournaise volcano at Reunion Island in the Indian Ocean.

6.6.4 GROUNDWATER FLOW PATTERN AT VULCANO (ITALY)

The La Fossa cone is an active stratovolcano located on Vulcano Island in the Aeolian archipelago in the Mediterranean Sea (southern Italy). Its self-potential anomaly (Revil et al., 2023; modified from Revil et al., 2008) exhibits a W shape. A 3-D resistivity tomogram was obtained by Revil et al. (2010b). The hydrothermal system is identified by very low values of the electrical resistivity (400 Ω -m). Inside the crater, it is possible to follow the plumbing system of the main fumarolic areas. On the flank of the edifice, a thick layer of tuff is also marked by very low resistivity values (in the range of 1–20 Ω -m) because of its composition in clays and zeolites. Inverse modelling shows that self-potential data can be used to determine the pattern of groundwater flow and potentially to assess water budget at the scale of the volcanic edifice. The distribution of the current density can then be translated into a distribution of Darcy velocity, which can be integrated to determine the amount of water upwelling through the central part of the volcano.

6.6.5 INVERSE MODELLING: THE CASE OF CERRO PRIETO (BAJA CALIFORNIA)

Revil et al. (2023) reported the interpretation of the 160 mV self-potential anomalies observed at the geothermal field of Cerro Prieto, Baja California, Mexico (Revil et al., 2023). The inversion of the self-potential and resistivity data allows observing a plume of hot groundwater rising to the ground surface in the central part of the investigated area and discharging to the ground surface in the south-western part. The temperature anomaly associated with the existence of this plume is independently observed by interpolating borehole temperature measurements (Revil et al., 2023). The analysis shown in Revil et al. (2023) summarizes the results obtained by Jardani and Revil (2009) using a stochastic approach.

In hydrothermal fields, the flow of the groundwater is responsible for both thermal and self-potential anomalies. Temperature is usually recorded in boreholes, while the self-potential field is recorded at the ground surface (Revil et al., 2023). The upwelling of a hydrothermal plume through a set of faults and permeable formations at the Cerro Prieto geothermal field in Baja California explains both the observed self-potential and the thermal anomalies (Revil et al., 2023).

6.6.6 UPFLOW OF THERMAL WATER ALONG A SHEAR FAULT (COLORADO)

A case study was done in the upper Arkansas Valley in Colorado, at the feet of Mount Princeton, which is part of the Collegial Peaks. Mount Princeton is the batholith of a former volcano and is composed of quartz monzonite. The Mount Princeton hot springs are partly associated with the upflow of thermal water along a major crustal shear fault. Such crustal faults are often associated with volcanism. The self-potential anomaly and shallow resistivity structure associated with an open conduit along this shear fault are shown in Revil et al. (2023; modified from Richards et al., 2010). The main objective here is to show how the flux of thermal water upwelling along the open conduit of the fault using the self-potential anomaly is quantified (Richards et al., 2010).

In order to quantify the flux of hot water, we need the geometrical model shown in Revil et al. (2023), which has been modified after Richards et al. (2010) with the resistivity values obtained from the resistivity tomogram shown in Revil et al. (2023). The geometry of the 2-D model is simplified in three units. Unit *U1* corresponds to the fault itself. Unit *U2* represents the quartz monzonite basement (mostly impervious), while *U3* denotes the shallow aquifer (made of granular sandy materials). The presence of this shallow aquifer is confirmed by a drill hole (labelled *MPG-5*). The presence of hot water (at a temperature of 59°C) was observed in well *MPG-5* at a depth of 40 m. The pH of the hot springs at Mount Princeton is comprised between 7.8 and 8.6. Then the Darcy velocity can be integrated along the dimension of the fault, characterized by the positive self-potential anomaly. This yields an integrated flux of 4,000 m³/day of hot water at 85°C (Richards et al., 2010). A similar approach was followed recently by Miller et al. (2018) for Mount Tongariro in New Zealand.

6.6.7 SELF-POTENTIAL ANOMALIES ASSOCIATED WITH THE ASCENT OF ACIDIC GROUNDWATER

The case of active volcanoes characterized by very acid fumarolic fluids (Revil et al., 2023) deserves a special attention. Two examples are provided in this section to illustrate the possible relationships between self-potential anomalies and hydrothermal fluids with very low pH values: (1) the case of La Soufriere volcano (Guadeloupe Island, France) and (2) the case of Papandayan volcanoes (Java Island, Indonesia). La Soufriere of Guadeloupe is characterized by the presence of a dome with fumarole activity, where boiling ponds exhibit extremely acidic fluids (minimal $\text{pH} \approx -0.8$) (Allard et al., 2014). Recent measurements also attest to values of pH below $\text{pH} = -1.0$ in the summit area (Inostroza et al., 2022). A multimethod profile, 2,430 m long, including self-potential measurements, soil temperature (at a depth of 30 cm), soil CO_2 concentration (at a depth of 30 cm), and electrical resistivity tomography, was performed in the southern flank of La Soufriere dome. The profile was located between ‘north of Piton Tarade’ on the western side of the profile and the ‘eastern flank of l’Echelle cone’ on the eastern side of the profile (labelled *NPT* and *EFE*, respectively, in Revil et al., 2023; modified from Brothelande et al., 2014).

On the western side of the profile, the self-potential data display three self-potential minima consistent with the presence of three gullies (Revil et al., 2023; modified from Brothelande et al., 2014). Along this profile, one area, <200 m long, was affected by high temperature and CO_2 concentration anomalies (close to 100°C and 100%, respectively). These anomalies are associated with a shallow conductive area located along the La Ty fault, a system of regional active tectonic faults crossing the volcanic dome (Revil et al., 2023; modified from Brothelande et al., 2014). While a positive anomaly is expected here, no self-potential anomaly is observed that could be associated with this upwelling of the hydrothermal fluids along the fault. The very low pH of the hydrothermal fluids is probably the cause of the absence of self-potential signature (for acidic fluids, the streaming potential coupling coefficient is close to zero).

Another example showing a clear impact of the pH of the upwelling hydrothermal fluids on the self-potential signals was observed at the Papandayan volcano (Byrdina et al., 2018). The upwelling of the acidic fluids is marked by the electrically conductive bodies observed on the 3-D resistivity tomogram. The self-potential measurements show a good correlation between the rising of hydrothermal fluids (Revil et al., 2023; modified from Byrdina et al., 2018) and the negative self-potential anomalies measured at the ground surface. The comparison between self-potential and soil pH provides an obvious positive correlation between these two parameters (Revil et al., 2023; modified from Byrdina et al., 2018). So it is therefore concluded that additional information such as the measurement of fluid or soil pH is critically needed to correctly interpret self-potential anomalies on stratovolcanoes.

6.6.8 TRANSIENT SELF-POTENTIAL ANOMALY

In volcanic environments, self-potential anomalies can fluctuate over time in a broad spectrum of frequencies from few milliseconds to several months or years, depending on the causative source of the signals. For instance, Kuwano et al. (2015) localized self-potential transient anomalies associated with very long period seismic pulses (Byrdina et al., 2003). Pride (1994) and Revil et al. (2015) show that the transient self-potential fluctuations are associated with any type of hydromechanical disturbances in partially water-saturated porous media. Revil et al. (2015) performed that the amplitude of the electric signals changed from the background level, indicating that a volume of water was flowing in the close vicinity of the well, producing fracking events at regular intervals. Such approach can be used to monitor hydromechanical events in a geothermal system. This implies the development of repeated surveys or the development of continuous monitoring networks on volcanoes (Yasukawa et al., 2005; Matsushima et al., 2017).

Repeated self-potential surveys performed at the Piton de la Fournaise volcano (Reunion Island, France) demonstrate how powerful the self-potential method is to detect areas of mechanical weakness before the occurrence of a major destructive event in the caldera of this shield volcano. Based

on a self-potential map performed over the entire cone of the Piton de la Fournaise volcano (Barde-Cabusson et al., 2012), a major self-potential positive anomaly located on the northern flank of the volcanic cone was chosen to install a reiteration profile, 440 m long, with measurement point each metre along the profile (Revil et al., 2023; modified after Chaput et al., 2019). The profile was reiterated twice, on 27 and 30 March 2007, a few days before a major collapse in the caldera that occurred in 5–7 April 2007 (Revil et al., 2023; modified after Chaput et al., 2019). The difference between these two self-potential reiterations displays three peaks corresponding to the maximum of difference (Revil et al., 2023; modified after Chaput et al., 2019). A different running average on the self-potential difference data allows better visualizing the self-potential changes. Two areas, *W* and *C*, correspond to a major difference in amplitude. The largest anomaly (*W*, 40 m of lateral extension) corresponds to the size of the so-called Soufriere pit crater before the 2007 collapse. Anomaly *C* corresponds precisely to the boundary of the caldera that collapsed in April 2007. Therefore, the fluctuation in the self-potential signals can be used to evidence areas of high permeability and mechanical weakness of a volcanic edifice. The mapping of these areas is very important in volcanology to assess, on a spatial point of view, structural weaknesses of a volcanic edifice, which can evolve during volcanic crisis in caldera, flank, or sector collapse.

This reiteration survey, performed each metre, demonstrates that interesting fluctuations in the self-potential signals can be measured only if one of the two electrodes is located over strategic areas (*W*, *C*, and *E* in this case). Two iterations of the profile are required. This result explains why permanent self-potential monitoring stations may remain blind to significant hydrothermal fluid disruptions if the electrodes are not correctly placed. Regarding permanent self-potential monitoring, it is interesting to mention the Izu-Oshima volcano (Japan), which has been continuously monitored from 2006 to 2014 (Matsushima et al., 2017). A decrease in the self-potential signal (~ 100 mV) has been observed at Izu-Oshima in the years following the last eruption in 1986–1987 (Matsushima et al., 2017). This decrease was explained by the reduction in gas emissions. Subsequent self-potential variations are mostly periodic, resulting mainly from changes in the water table level due to precipitation. To better understand future self-potential precursor events, it will be necessary to measure the self-potential signal over one or several eruptive cycles.

6.7 SELF-POTENTIAL (SP) ANOMALY OF SULPHIDE ORE BODY, SARIYER AREA, ISTANBUL, TURKEY: A CASE STUDY

The self-potential method has a wide range in different applications in the exploration of geophysics, hydrogeophysics, and environmental problems and mineral exploration, such as metallic sulphides, magnetite, graphite, and uranium. Several assessable elucidation approaches of the SP data over the buried geologic structures have been established. Figure 6.5 (Essa and Munschy, 2019) displays the self-potential anomaly of a sulphide ore body in the Sariyer area, which is located about 18 km north of Istanbul, Turkey, and characterized by an outcropping of andesite, pyrite veins, and cupriferous waters.

The area of this investigation is characterized by a steep surface gradient. In 1951, the sulphide ore bodies had been explored by utilizing geophysical techniques, where it was found to lie under unmineralized schist or alluvium with a depth of 23 m and elongated as a spheroid dimension (Essa, 2011). This profile has a length of 160 m and has been subjected to many interpretation methods to estimate the sulphide ore body parameters (K is the amplitude factor, z is the depth, θ is the polarization angle, d is the origin location, and q is the shape). The estimated results are displayed in Table 6.2. The estimated parameters of this source by exploiting these methods have a good covenant together.

6.8 TELLURIC CURRENTS

Ultraviolet radiation from the sun ionizes molecules of air in the thin upper atmosphere of the Earth. The ions accumulate in several layers, forming the ionosphere at altitudes between about 80 km and 1,500 km above the Earth's surface. Electric currents in the ionosphere arise from systematic

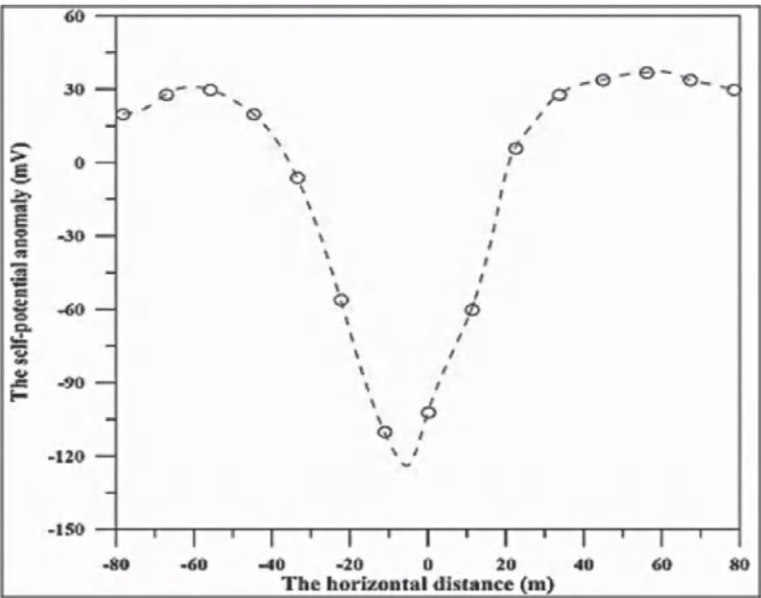


FIGURE 6.5 The self-potential anomaly of a sulphide ore body in the Sariyer area, Turkey.

TABLE 6.2
Results Obtained for Interpreting Self-Potential Anomaly of a Sulphide Ore Body in the Sariyer Area, Turkey

Parameters	Abdelrahman et al. (2004) Method	Asfahani and Tlas (2016) Method	Biswas and Sharma (2015) Method	Essa (2011) Method
K (mV × m ²)	3,245.0	4,695.6	2,855.4	3,447.1
z (m)	24.5	22.6	28.0	25.4
θ (°)	−51.2	−82.9	−70.3	−58.9
d (m)	—	—	−2.5	−0.5
q (dimensionless)	1.5	1.0	1.5	1.5

Source: Essa and Munsch (2019).

motions of the ions, which are affected by various factors, such as the daily and monthly tides, seasonal variations in insolation, and the periodic fluctuation in ionization related to the 11-year sunspot cycle. The currents produce varying magnetic fields with the same frequencies, which are observed at the surface of the Earth and can be analyzed from long-term continuous records of the geomagnetic field. The ionospheric effects show up in the energy spectrum of the geomagnetic field as distinct peaks representing periods that range from fractions of a second (geomagnetic pulsations) to several years (Figure 6.6) (Serson, 1973).

The magnetic fields induce fluctuating electric currents, called telluric currents, that flow in horizontal layers in the crust and mantle. The current pattern consists of several huge whorls, thousands of kilometres across, which remain fixed with respect to the sun and thus move around the Earth as it rotates.

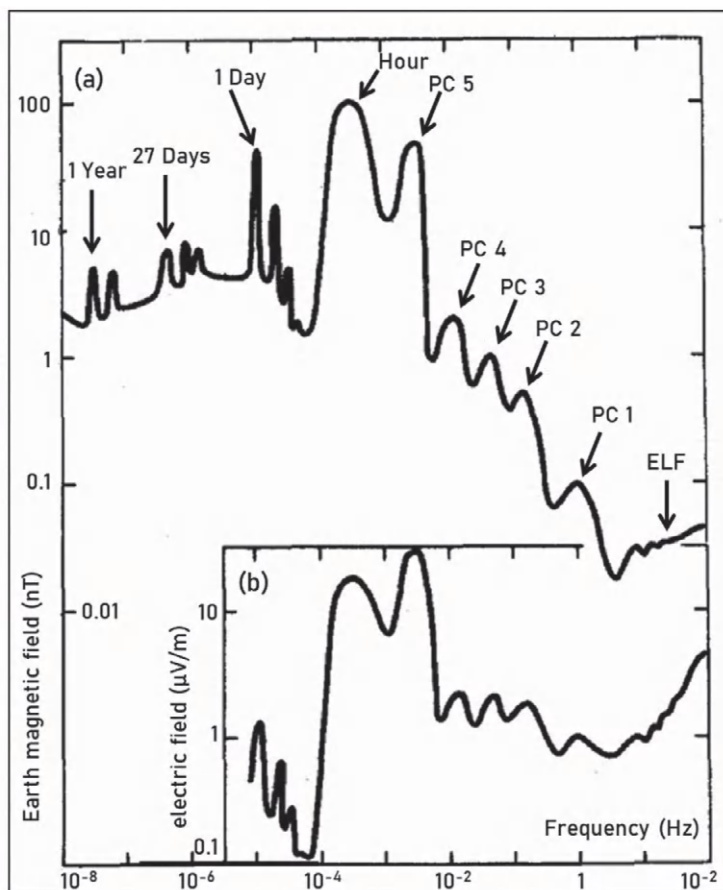


FIGURE 6.6 (a) The frequency spectrum of natural variations in the horizontal intensity of the geomagnetic field, and (b) the corresponding spectrum of induced electric field fluctuations, computed for a model Earth with uniform resistivity of 20 $\Omega\text{-m}$.

The distribution of telluric current density depends on the variation of resistivity in the horizontal conducting layers. At shallow crustal depths, the lines of current flow are disturbed by subsurface structures, which cause contrasts in resistivity. These could arise from geological structures or the presence of mineralized zones. Consider, for example, a buried anticline which has a highly resistive rock (such as granite) as its core and is overlain by a conducting layer of porous sedimentary rocks saturated with groundwater. The horizontal flow of telluric current across the anticline chooses the less-resistive path through the conducting sediments. The current lines bunch together over the axis of the anticline, increasing the horizontal current density (Figure 6.7). The equipotential surfaces normal to the current lines intersect the ground surface, where potential differences can be measured with a high-impedance voltmeter.

The field equipment for measuring telluric current density is simple. The sensors are a pair of nonpolarizable electrodes with a fixed separation L of the order of 10–100 m. The potential difference V between the electrodes is measured with a high-impedance voltmeter. The electric field E at a point midway between the electrodes can be assumed to be V/L . Using Ohm's law, and assuming that the telluric current flows in conducting rock layer with resistivity ρ_1 , the telluric current density J at each measurement station along a profile is given by

$$J = \frac{V}{\rho_1 L}$$

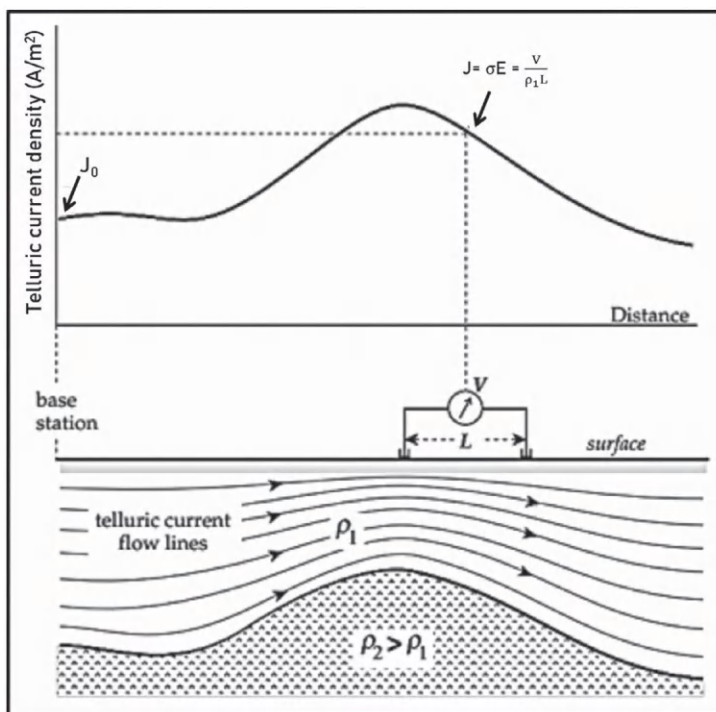


FIGURE 6.7 Telluric current lines are deflected by changes in thickness of a conducting layer over a more resistive structure (bottom). The telluric current density (top) is obtained from the voltage measured between a pair of fixed-separation electrodes at the surface.

Source: Robinson and Coruh (1988).

The deflection of telluric current by a resistive subsurface structure as shown in Figure 6.7 is greatly idealized. It assumes an infinite resistivity ρ_2 in the core of the anticline. In practice, the current is not completely diverted through the better-conducting layer; part flows through the more-resistive layer as well. Thus, we cannot assume that the resistivity ρ_1 corresponds to the good conductor. Rather, it represents some undefined mixture of the values ρ_1 and ρ_2 . It is not the true resistivity of either layer but the apparent resistivity of the measurement.

Finally, the three potential methods (gravity, magnetic, and self-potential) have been used to evaluate the source parameters but are suffering of ill-posedness and non-uniqueness in finding a global solution (Tarantola, 2004). The usage of simple geometrical structures in gravity, magnetic, and self-potential inversion helps in overcoming some of these limitations, gives an optimal fit for the buried structures, and plays a vigorous role in solving many investigation problems.

REFERENCES

- Abdelrahman, E.M., Saber, H.S., Essa, K.S. and Fouda, M.A., 2004. A least-squares approach to depth determination from numerical horizontal self-potential gradients. *Pure and Applied Geophysics*, vol. 161, pp. 399–411.
- Ahmad, M.U., 1969. A laboratory study of streaming potentials. *Geophysical Prospecting*, vol. 12, no. 1, pp. 49–64.
- Aizawa, K., 2000. Classification of self-potential anomalies on volcanoes and possible interpretations for their subsurface structure. *Journal of Volcanology and Geothermal Research*, vol. 175, pp. 253–268.
- Aizawa, K., 2004. A large self-potential anomaly and its changes on the quiet Mt. Fuji, Japan. *Geophysical Research Letters*, vol. 31, no. 5.

- Aizawa, K., Yoshimura, R., Oshiman, N., Yamazaki, K., Uto, T., Ogawa, Y. and Hurst, W.A., 2005. Hydrothermal system beneath Mt. Fuji volcano inferred from magnetotellurics and electric self-potential. *Earth and Planetary Science Letters*, vol. 235, no. 1–2, pp. 343–355.
- Allard, P., Aiuppa, A., Beauducel, F., Gaudin, D., Di Napoli, R., Calabrese, S. and Tamburello, G., 2014. Steam and gas emission rate from La Soufriere volcano, Guadeloupe (Lesser Antilles): Implications for the magmatic supply during degassing unrest. *Chemical Geology*, vol. 384, pp. 76–93.
- Anderson, L.A. and Johnson, C.R., 1976. Application of the self-potential method to geothermal exploration in Long Valley, California. *Journal of Geophysical Research*, vol. 81, pp. 1527–1532.
- Asfahani, J. and Tlas, M., 2016. Interpretation of self-potential anomalies by developing an approach based on linear optimization. *Geosciences and Engineering*, vol. 5, pp. 7–21.
- Atekwana, E.A. and Atekwana, E.A., 2010. Geophysical of microbial activity at hydrocarbon contaminated Sites: A Review. *Surveys in Geophysics*, vol. 31, pp. 247–283.
- Aubert, M. and Dana, I., 1994. Interpretation of the self-potential radial profiles in volcanology: Possibilities of the SP method for the monitoring of the active volcanoes. *Bulletin de la Societe Geologique de France*, vol. 2, pp. 113–122.
- Barde-Cabusson, S., Finizola, A., Peltier, A., Chaput, M., Taquet, N., Dumont, S., Duputel, Z., Guy, A., Mathieu, L., Saumet, S., Sorbadere, F. and Vieille, M., 2012. Structural control of collapse events inferred by self-potential mapping on the Piton de la Fournaise volcano (La Reunion Island). *Journal of Volcanology and Geothermal Research*, no. 209–210, pp. 9–18.
- Barde-Cabusson, S., Finizola, A., Revil, A., Ricci, T., Piscitelli, S., Rizzo, E. and Villeneuve, N., 2009b. New geological insights and structural control on fluid circulation in La Fossa cone (Volcano, Aeolian Islands, Italy). *Journal of Volcanology and Geothermal Research*, vol. 185, no. 3, pp. 231–245.
- Barde-Cabusson, S., Levieux, G., Lenat, J.F., Finizola, A., Revil, A., Chaput, M. and Vieille, M., 2009a. Transient self-potential anomalies associated with recent lava flows at Piton de la Fournaise volcano (Reunion Island, Indian Ocean). *Journal of Volcanology and Geothermal Research*, vol. 187, no. 3–4, pp. 158–166.
- Bennati, L., Finizola, A., Walker, J., Lopez, D., Higuiera-Diaz, C., Schutze, C., Barahona, F., Cartagena, R., Conde, V., Funes, R. and Rios, C., 2011. Fluid circulation in a complex volcano-tectonic setting, inferred from self-potential and soil CO₂ flux surveys: The Santa María-Cerro Quemado-Zunil volcanoes and Xela caldera (northwestern Guatemala). *Journal of Volcanology and Geothermal Research*, vol. 199, pp. 216–229.
- Biswas, A. and Sharma, S.P., 2015. Interpretation of self-potential anomaly over idealized body and analysis of ambiguity using very fast simulated annealing global optimization. *Near Surface Geophysics*, vol. 13, pp. 179–195.
- Bogdan, Z., Mendecki, M.J., Zuberek, W.M. and Robak, M., 2012. Application of self-potential method in the area contaminated with oil derivatives. *Acta Geodynamica et Geomaterialia*, vol. 9, no. 2 (166), pp. 179–189.
- Bogoslovsky, V.A. and Ogilvy, A.A., 1973. Deformations of natural electric fields near drainage structures. *Geophysical Prospecting*, vol. 21, pp. 716–723.
- Bogoslovsky, V.V. and Ogilvy, A.A., 1970a. Applications of geophysical methods for studying the technical status of Earth Dams. *Geophysical Prospecting*, vol. 18, pp. 758–773.
- Bogoslovsky, V.V. and Ogilvy, A.A., 1970b. Natural potential anomalies as a qualitative index of the rate of seepage from water reservoir. *Geophysics*, vol. 18, no. 2, pp. 261–268.
- Boleve, A., Janod, F., Revil, A., Lafon, A. and Fry, J.J., 2011. Localization and quantification of leakages in dams using time-lapse self-potential measurements associated with salt tracer injection. *Journal of Hydrology*, vol. 403, nos. 3–4, pp. 242–252.
- Brothelande, E., Finizola, A., Peltier, A., Delcher, E., Komorowski, J.C., Di Gangi, F. and Legendre, Y., 2014. Fluid circulation pattern inside La Soufriere volcano (Guadeloupe) inferred from combined electrical resistivity tomography, self-potential, soil temperature and diffuse degassing measurements. *Journal of Volcanology and Geothermal Research*, vol. 288, pp. 105–122.
- Butler, D.K., Llopis, J.I., Dobecki, T.L., Corwin, R.F., Wilt, B.J. and Olhoeft, G., 1990. Comprehensive geophysics investigation of an existing dam foundation: Engineering geophysics research and development, geophysics. *The Leading Edge of Exploration Geotechnical Applications*, vol. 5, part 2, pp. 44–53.
- Byrdina, S., Friedel, S., Wassermann, J. and Zlotnicki, J., 2003. Self-potential variations associated with ultra-long-period seismic signals at Merapi volcano. *Geophysical Research Letters*, vol. 30, no. 22, pp. 2156.
- Byrdina, S., Grandis, H., Sumintadireja, P., Caudron, C., Syahbana, D.K., Naffrechoux, E., ..., Vandemeulebrouck, J., 2018. Structure of the acid hydrothermal system of Papandayan volcano, Indonesia, investigated by geophysical methods. *Journal of Volcanology and Geothermal Research*, vol. 358, pp. 77–86.

- Chaput, M., Finizola, A., Peltier, A., Villeneuve, N., Crovisier, M. and Barde-Cabusson, S., 2019. Where does a volcano break? Using self-potential reiteration to forecast the precise location of major destructive events on active volcanoes: The case study of the Piton de la Fournaise 2007 caldera collapse. *Volcanica*, vol. 2, no. 2, pp. 151–159.
- Coleman, A.R., 1991. The use of the self-potential method in the delineation of a reclaimed landfill site. Proceedings of the Conference on Planning and Engineering of Landfills, 10–11 July, University of Bartington, The Midlands Geotechnical Society, Birmingham, West Midlands, B15 2TT, GB.
- Corry, C.E., 1985. Spontaneous polarization associated with porphyry sulphide mineralization. *Geophysics*, vol. 50, pp. 1020–1034.
- Corwin, A.F., Morrison, H.F., Diaz, C.S. and Rodriguez, B.J., 1979. Self-potential studies at the Cerro Prieto geothermal field. Proceedings, First Symposium on the Cerro Prieto Geothermal Field, Baja California, Mexico, September 1978, 7098, Lawrence Berkeley Laboratory, LBL, Berkeley, pp. 204–210.
- Corwin, R.F., 1990. The self-potential method for environmental and engineering applications. In: Ward, S.H. (ed), *Geotechnical and Environmental Geophysics, vol. 1: Review and Tutorial*. Tulsa, OK: Society of Exploration Geophysicists, pp. 127–145.
- Corwin, R.F. and Hoover, D.B., 1979. The self-potential method in geothermal exploration. *Geophysics*, vol. 44, no. 2, p. 226.
- Davenport, G.C., Hadley, L.M. and Randall, J.A., 1978. The use of seismic refraction and self-potential surveys to evaluate existing embankments. Available at www.heritagegeophysics.com/papers/dave6.htm, July 2006.
- Essa, K.S., 2011. A new algorithm for gravity or self-potential data interpretation. *Journal of Geophysics and Engineering*, vol. 8, pp. 434–446.
- Essa, K.S. and Munsch, M., 2019. *Introductory Chapter: Mineral Exploration from the Point of View of Geophysicists*. Minerals (Book), DOI: 10.5772/intechopen.84830.
- Finizola, A., Aubert, M., Revil, A., Schutze, C. and Sortino, F., 2009. Importance of structural history in the summit area of Stromboli during the 2002–2003 eruptive crisis inferred from temperature, soil CO₂, self-potential, and electrical resistivity tomography. *Journal of Volcanology and Geothermal Research*, vol. 183, pp. 213–227.
- Finizola, A., Lenat, J.-F., Macedo, O., Ramos, D., Thouret, J.-C. and Sortino, F., 2004. Fluid circulation and structural discontinuities inside Misti volcano (Peru) inferred from self-potential measurements. *Journal of Volcanology and Geothermal Research*, vol. 135, no. 4, pp. 343–360.
- Finizola, A., Sortino, F., Lenat, J.-F., Aubert, M., Ripepe, M. and Valenza, M., 2003. The summit hydrothermal system of Stromboli: New insights from self-potential, temperature, CO₂ and fumarolic fluids measurements, with structural and monitoring implications. *Bulletin of Volcanology*, vol. 65, pp. 486–504.
- Finizola, A., Sortino, F., Lenat, J.-F. and Valenza, M., 2002. Fluid circulation at Stromboli volcano (Aeolian Islands, Italy) from self-potential and soil gas surveys. *Journal of Volcanology and Geothermal Research*, vol. 116, no. 1–2, pp. 1–18.
- Fournier, C., 1989. Spontaneous potentials and resistivity surveys applied to hydrogeology in a volcanic area: Case history of the Chaîne des Puys (Puy-de-Dôme, France). *Geophysical Prospecting*, vol. 37, pp. 647–668.
- Fox, R.W., 1830. On the electromagnetic properties of metalliferous veins in the mines of Cornwall. *Philosophical Transactions of the Royal Society*, vol. 130, pp. 399–414.
- Fraseri, A., 2002. Relations between the hydrocarbon migration through chimney and the electric self-potential field. *Journal of the Balkan*, no. 7, pp. 311–316.
- Gex, P., 1980. Electrofiltration phenomena associated with several dam sites. *Bulletin de la Société Vaudoise des Sciences Naturelles*, vol. 357, no. 75, pp. 39–50.
- Gonzales, K., Finizola, A., Lenat, J.-F., Macedo, O., Ramos, D., Thouret, J.-C., Fournier, N., Cruz, V. and Pistre, K., 2014. Asymmetrical structure, hydrothermal system and edifice stability: The case of Ubinas volcano, Peru, revealed by geophysical surveys. *Journal of Volcanology and Geothermal Research*, vol. 276, pp. 132–144.
- Gresse, M., Uyeshima, M., Koyama, T., Hase, H., Aizawa, K., Yamaya, Y. and Hata, M., 2021. Hydrothermal and magmatic system of a volcanic island inferred from magnetotellurics, seismicity, self-potential, and thermal image: An example of Miyakejima (Japan). *Journal of Geophysical Research: Solid Earth*, vol. 126, no. 6.
- Haas, A.K., Revil, A., Karaoulis, M., Frash, L., Hampton, J., Gutierrez, M. and Mooney, M., 2013. Electrical potential source localization reveals a borehole leak during hydraulic fracturing. *Geophysics*, vol. 78, no. 2, pp. D93–D113.

- Hase, H., Hashimoto, T., Sakanaka, S., Kanda, W. and Tanaka, Y., 2005. Hydrothermal system beneath Aso volcano as inferred from self-potential mapping and resistivity structure. *Journal of Volcanology and Geothermal Research*, vol. 143, pp. 259–277.
- Howells, K. and Fox, D., 1998. Geophysical methods for detecting shallow sulphide mineralization in the Halifax Formation, Nova Scotia: A reconnaissance study. *Atlantic Geology*, vol. 34, pp. 211–227.
- Huenges, E., 2010. *Geothermal Energy Systems: Exploration, Development, and Utilization*. Hoboken, NJ: Wiley Publisher, p. 487.
- Inostroza, M., Moune, S., Moretti, R., Robert, V., Bonifacie, M., Chilin-Eusebe, E. and Burckel, P., 2022. Monitoring hydrothermal activity using major and trace elements in low-temperature fumarolic condensates: The case of La Soufriere de Guadeloupe. *Volcano Geosciences*, vol. 12, no. 7, p. 267.
- Ishido, T., Kikuchi, T., Matsushima, N., Yano, Y., Nakao, S., Sugihara, M., Tosha, T., Takakura, S. and Ogawa, Y., 1997. Repeated self-potential profiling of Izu-Oshima volcano, Japan. *Journal of Geomagnetism and Geoelectricity*, vol. 49, no. 11–12, pp. 1267–1278.
- Jardani, A. and Revil, A., 2009. Stochastic joint inversion of temperature and self-potential data. *Geophysical Journal International*, vol. 179, no. 1, pp. 640–654.
- Jolie, E., Scott, S., Faulds, J., Chambefort, I., Axelsson, G., Gutierrez-Negrín, L.C. and Zemedkun, M.T., 2021. Geological controls on geothermal resources for power generation. *Nature Reviews Earth and Environment*, vol. 2, no. 5, pp. 324–339.
- Keller, G.V., 1981. Exploration for geothermal energy. In: Fitch, A.A. (ed), *Developments in Geophysical Exploration: Methods—2*. Basel: MDPI, pp. 107–150.
- Kuwano, O., Yoshida, S., Nakatani, M. and Uyeshima, M., 2015. Origin of transient self-potential signals associated with very long period seismic pulses observed during the 2000 activity of Miyakejima volcano. *Journal of Geophysical Research: Solid Earth*, vol. 120, no. 5, pp. 3544–3565.
- Lowrie, W., 2007. *Fundamentals of Geophysics*. 2nd edn. New York: United States of America by Cambridge University Press, p. 375.
- Mahardika, H., Revil, A. and Jardani, A., 2012. Waveform joint inversion of seismograms and electrograms for moment tensor characterization of fracking events. *Geophysics*, vol. 77, no. 5, pp. ID23–ID39.
- Malama, B., Kuhlman, K.L. and Revil, A., 2009a. Theory of transient streaming potentials associated with axial-symmetric flow in unconfined aquifers. *Geophysical Journal International*, vol. 179, pp. 990–1003.
- Malama, B., Revil, A. and Kuhlman, K.L., 2009b. A semi-analytical solution for transient streaming potentials associated with confined aquifer pumping tests. *Geophysical Journal International*, vol. 176, pp. 1007–1016.
- Matsushima, N., Nishi, Y., Onizawa, S.Y., Takakura, S., Hase, H. and Ishido, T., 2017. Self-potential characteristics of the dormant period of Izu-Oshima volcano. *Bulletin of Volcanology*, vol. 79, no. 12, p. 86.
- Meiser, P., 1962. A method for quantitative interpretation of self-potential measurements. Presented at the Twentieth Meeting of the European Association of Exploration Geophysicists, held in the Hague, 31 May to 3 June, 1961. First published: June 1962, Geophysical prospecting, DOI: 10.1111/j.1365-2478.1962.tb02009.
- Miller, C.A., Kang, S.G., Fournier, D. and Hill, G., 2018. Distribution of and condensate in a hydrothermal system: Insights from self-potential inversion at Mount Tongariro, New Zealand. *Geophysical Research Letters*, vol. 45, pp. 8190–8198.
- Moore, J.R. and Glaser, S.D., 2007. Self-potential observations during hydraulic fracturing. *Journal of Geophysical Research*, vol. 112, pp. B02204.
- Naskar, D.C. and Kulkarni, A.V., 2019. Geophysical investigation for copper and gold mineralization around Dhani-Basri and Todi ka Bas area, Dausa and Alwar districts, Rajasthan. *Journal of Indian Geophysical Union*, vol. 23, no. 2, pp. 120–133.
- Naskar, D.C., Kulkarni, A.V. and Lakshmana, M., 2018. Geophysical investigation for base metal mineralization in Karoi-Rajpura area, Bhilwara district, Rajasthan. *Journal of Indian Geophysical Union*, vol. 22, no. 4, pp. 379–388.
- Nayak, P.N., 1981. Electromechanical potential in surveys for sulphides. *Geoexploration*, vol. 18, pp. 311–320.
- Nordiana, M.M., Olugbenga, A.T., Saharudin, M.A., Nabila, S. and El Hidayah Ismail, N., 2017. The application of 2-D resistivity and Self Potential (SP) methods in determining the water flow. IOP Conference Series: Journal of Physics: Conference Series, vol. 995, p. 012077, IOP Publishing, DOI: 10.1088/1742-6596/995/1/012077, p. 9.
- Nyquist, J.E. and Corry, C.E., 2002. Self-potential: The ugly duckling of environmental geophysics. *The Leading Edge*, vol. 21, no. 5, pp. 446–451.
- Panthulu, V., Krishnaiah, C. and Shirke, J.M., 2001. Detection of seepage paths in earth dams using self-potential and resistivity methods. *Engineering Geology*, vol. 59, pp. 281–295.

- Park, S., Wolf, L.W., Lee, M. and Saunders, J., 2004. Self-potential and geochemical measurements of microbially mediated bacterial sulphate reduction in saturated sediments. American Geophysical Union, Fall Meeting 2004, abstract #B53A-0983.
- Petrovski, A., 1928. The problem of a hidden polarized sphere. *Philosophical Magazine*, vol. 5, pp. 334–358.
- Poldini, E., 1938. Geophysical exploration by spontaneous polarization methods. *Mineralogical Magazine*, vol. 59, pp. 278–282, 347–352.
- Poldini, E., 1939. Geophysical exploration by spontaneous polarization methods. *Mineralogical Magazine*, vol. 60, pp. 22–27, 90–94.
- Pride, S., 1994. Governing equations for the coupled electromagnetics and acoustics of porous media. *Physical Review B*, vol. 50, pp. 15678–15696.
- Revil, A., Fan, Q., Cary, L., Finizola, A. and Trolard, F., 2005a. Self-potential signals associated with preferential groundwater flow pathways in a buried paleo-channel. *Geophysical Research Letters*, vol. 32, p. L07401.
- Revil, A., Finizola, A. and Gresse, M., 2023. Self-potential as a tool to assess groundwater flow in hydrothermal systems: A review. *Journal of Volcanology and Geothermal Research*, vol. 437, 107788.
- Revil, A., Finizola, A., Piscitelli, A., Rizzo, E., Ricci, T., Crespy, A., Angeletti, B., Balasco, M., Barde, C.S., Bennati, L., Boleve, A., Byrdina, S., Carzaniga, N., Di Gangi, F., Morin, J., Perrone, A., Rossi, M., Roulleau, E. and Suski, B., 2008. Inner structure of La Fossa di Vulcano (Vulcano Island, southern Tyrrhenian Sea, Italy) revealed by high resolution electric resistivity tomography coupled with self-potential, temperature, and soil CO₂ diffuse degassing measurements. *Journal of Geophysical Research*, vol. 113, pp. B07207.
- Revil, A., Finizola, A., Sortino, F. and Ripepe, M., 2004. Geophysical investigations at Stromboli volcano, Italy: Implications for ground water flow and paroxysmal activity. *Geophysical Journal International*, vol. 157, pp. 426–440.
- Revil, A., Johnson, T.C. and Finizola, A., 2010b. Three-dimensional resistivity tomography of Vulcan's forge, Vulcano Island, southern Italy. *Geophysical Research Letters*, vol. 37, p. L15308.
- Revil, A., Leroy, P. and Titov, K., 2005b. Characterization of transport properties of argillaceous sediments. Application to the Callovo-Oxfordian argillite. *Journal of Geophysical Research*, vol. 110, p. B06202.
- Revil, A., Mao, D., Haas, A.K., Karaoulis, M. and Frash, L., 2015. Passive electrical monitoring and localization of leakages in wells using electrography. *Journal of Hydrology*, vol. 521, pp. 286–301.
- Revil, A. and Pezard, P.A., 1998. Streaming electrical potential anomaly along faults in geothermal areas. *Geophysical Research Letters*, vol. 25, no. 16, pp. 3197–3200.
- Revil, A., Pezard, P.A. and Glover, P.W.J., 1999a. Streaming potential in porous media. 1. Theory of the zeta-potential. *Journal of Geophysical Research*, vol. 104, no. B9, pp. 20021–20031.
- Revil, A., Schwaeger, H., Cathles, L.M. and Manhardt, P., 1999b. Streaming potential in porous media: 2. Theory and application to geothermal systems. *Journal of Geophysical Research*, vol. 104, no. B9, pp. 20033–20048.
- Reynolds, J.M., 1997. *An Introduction to Applied and Environmental Geophysics*. New York City: John Wiley and Sons Inc., p. 796.
- Richards, K., Revil, A., Jardani, A., Henderson, F., Batzle, M. and Haas, A., 2010. Pattern of shallow ground water flow at Mount Princeton Hot Springs, Colorado, using geoelectrical methods. *Journal of Volcanology and Geothermal Research*, vol. 198, pp. 217–232.
- Robinson, E.S. and Coruh, C., 1988. *Basic Exploration Geophysics*. New York: John Wiley and Sons.
- Sato, M. and Mooney, H.M., 1960. The electro-chemical mechanism of sulphide self-potentials. *Geophysics*, vol. 25, no. 1, pp. 226–249.
- Semenov, A.S., 1980. *Electric Prospecting by Self-Potential Method*, 4th ed.; revised and supplemented. Leningrad, Russia: Nedra.
- Serson, P.H., 1973. Instrumentation for induction studies on land. *Physics of the Earth and Planetary Interiors*, vol. 7, pp. 313–322.
- Sill, W.R., 1983. Self-potential modeling from primary flows. *Geophysics*, vol. 48, pp. 76–86.
- Sill, W.R. and Bodell, J., 1977. Thermal gradients and heat flow at Roosevelt Hot Springs: University of Utah. *Department of Geology and Geophysics Report*, vol. 77, no. 3, p. 46.
- Soueid Ahmed, A., Jardani, A., Revil, A. and Dupont, J.P., 2016. Specific storage and hydraulic conductivity tomography through the joint inversion of hydraulic heads and self-potential data. *Advances in Water Resources*, vol. 89, pp. 80–90.
- Spies, B.R. and Frischknecht, F.C., 1992. Electromagnetic sounding. In: Nabighian, M.N. (ed), *Electromagnetic Methods in Applied Geophysics*, vol. 2. Houston, TX and Tulsa, OK: SEG, pp. 285–426.

- Stern, W., 1945. Relation between spontaneous polarization curves and depth, size, and dip of ore bodies. *Transactions of the American Institute of Mining and Metallurgical Engineers*, vol. 164, pp. 189–196.
- Suski, B., Revil, A., Titov, K., Konosavsky, P., Dages, C., Voltz, M. and Huttel, O., 2006. Monitoring of an infiltration experiment using the self-potential method. *Water Resources Research*, vol. 42, p. W08418.
- Suski, B., Rizzo, E. and Revil, A., 2004. A sandbox experiment of self-potential signals associated with a pumping test. *Vadose Zone Journal*, vol. 3, pp. 1193–1199.
- Tarantola, A., 2004. *Inverse Problem Theory and Methods for Model Parameter Estimation*. 1st edn. Philadelphia: Society for Industrial and Applied Mathematics (SIAM), p. 352.
- Teh Saufia, A.H.A., Saad, R., Nordiana, M.M., Azwin, I.N., Bery, A.A., Noer, E.H.I., Nisa, A., Mark, J., Kiu, Y.C. and Ragu, R.R. 2013. Geoelectrical survey for engineering investigation. *Open Journal of Geology*, no. 3, pp. 46–49.
- Villasante-Marcos, V., Finizola, A., Abella, R., Barde-Cabusson, S., Blanco, M.J., Brenes, B., Cabrera, V., Casas, B., De Agustin, P., Di Gangi, F., Dominguez, I., Garcia, O., Gomis, A., Guzman, J., Iribarren, I., Levieux, G., Lopez, C., Luengo, N., Martín, I., Moreno, M., Meletlidis, S., Morin, J., Moure, D., Pereda, J., Ricci, T., Romero, E., Schutze, C., Suski-Ricci, B., Torres, P. and Trigo, P., 2014. Hydrothermal system of Central Tenerife Volcanic Complex, Canary Islands (Spain), inferred from self-potential measurements. *Journal of Volcanology and Geothermal Research*, vol. 272, pp. 59–77.
- Vozoff, K., 1992. The magnetotelluric method. In: Nabighian, M.N. (ed), *Electromagnetic Methods in Applied Geophysics*, vol. 2. Houston, TX and Tulsa, OK: SEG, pp. 641–712.
- Wynn, J.C. and Sherwood, S.I., 1984. The Self-Potential (SP) Method: An Inexpensive Reconnaissance and Archaeological Mapping Tool. *Journal of Field Archaeology*, vol. 11, no. 2, pp. 195–204.
- Yasukawa, K., Ishido, T. and Suzuki, I., 2005. Geothermal reservoir monitoring by continuous self-potential measurements, Mori geothermal field, Japan. *Geothermics*, vol. 34, no. 5, pp. 551–567.
- Zlotnicki, J. and Nishida, S., 2003. Review on morphological insights of self-potential anomalies on volcanoes. *Surveys in Geophysics*, vol. 24, no. 4, pp. 291–338.
- Zohdy, A.A.R., Anderson, L.A. and Mufler, L.J.P., 1973. Resistivity, self-potential and induced-polarization surveys of a vapour dominated geothermal system. *Geophysics*, vol. 38, pp. 1130–1144.

7 Induced Polarization (IP) Method

7.1 INTRODUCTION

The Schlumberger brothers (pioneers in geophysical exploration) recognized the phenomenon of IP in the 1950s. *Induced polarization* (IP) was developed to aid the exploration of disseminated electronically conducting mineralization that cannot be detected by resistivity and electromagnetic (EM) techniques.

The use of the induced electrical polarization (IP) technique over the past 30 years has proven to be one of the most successful geophysical methods in providing in situ information about rock mineralogy, especially in the search for disseminated minerals with electronic conductivity. For this case, the technique is applied to quite a new field which concerns materials that do not contain conductive minerals but rather clay minerals. The method has been used, though rarely, in the fields of hydrogeology (Vacquier et al., 1957), (Marshall and Madden, 1959), in oil and gas field exploration (Sternberg and Oehler, 1990), and in environmental studies, such as mapping of polluted land areas (Towel et al., 1985). However, it is worth noting that previous IP studies of nonmetallic Earth materials have focused on clay mineral soils, sandy and shaly sediments containing clay minerals (Klein and Sill, 1982).

All metallic minerals (lithium, beryllium, lead, zinc, copper, gold, silver, iron, manganese, uranium, thorium, molybdenum, tungsten, tin, vanadium, rare Earth elements, and tantalum), graphite, and some oxides, including magnetite, are good for IP target. IP has become an indispensable tool in the search for base metal (Cu, Pb, and Zn) and Au mineralization—volcanogenic massive sulphide.

In India, the first IP surveys were conducted in 1966 under the project ‘Operation Hard Rock’, which later formed the Airborne Mineral Survey Exploration (AMSE) wing of the Geological Survey of India (GSI). However, the wide application of the IP method was only started in India by GSI from 1970 after the import of the equipment from the Canadian Mining Geophysical Companies.

GSI has been using Robertson Geologging technology throughout India for over 30 years. The mineral exploration and mining history of India is closely related to the establishment of the Geological Survey of India (GSI). Mineral resource assessment projects include ferrous minerals, base metals, precious metals and minerals, nonferrous and strategic minerals, and industrial and fertilizer. Lately, GSI has taken the important policy decision of extending its regional work to a detailed exploration stage in potential development areas, ensuring that comprehensively explored areas can be offered to the stakeholders for exploitation. GSI is using a range of Robertson Geo tools (UK Robertson Geologging Ltd.; Americas Robertson Geologging [USA] Inc.; Asia Robertson Geologging [Asia] Inc.) for its exploration projects, including IP, magnetic susceptibility, spectral gamma, resistivity, density, neutron, etc. It also has a logging depth deployment capability of over 1,800 m. Recent projects have included the economic feasibility of potash mining in the Hanumangarh District of Rajasthan. After preliminary exploration, GSI has found potash reserves of 133 million tonnes in one area of the district (GSI Annual Report). Another long-time Robertson Geo client, Mineral Exploration Corporation Limited (MECL), will now undertake a more detailed study of the reserves.

7.2 BASIC PRINCIPLE (POLARIZATION OF THE EARTH)

Polarization is a geophysical phenomenon which measures the slow decay of voltage in the ground after the cessation of an excitation current pulse (time domain method) or low-frequency variation of the resistivity of the Earth (frequency domain method) (Sumner, 1976). In simple terms, the IP

effect reflects the degree to which the subsurface is able to store electric charge, analogous to a leaky capacitor. It occurs when an electric current passes through a rock/soil. If commutated direct current is used in a four-electrode resistivity survey, the sequence of positive and negative flow may be interspersed with periods when the current is off. The inducing current then has a box-like appearance (Figure 7.1a). When the current is interrupted, the voltage across the potential electrodes does not drop immediately to zero. After an initial abrupt drop to a fraction of its steady-state value, it decays slowly for several seconds (Figure 7.1b). Conversely, when the current is switched on, the potential rises suddenly at first and then gradually approaches the steady-state value. The slow decay and growth of part of the signal are due to induced polarization, which results from two similar effects related to the rock structure: membrane polarization and electrode polarization.

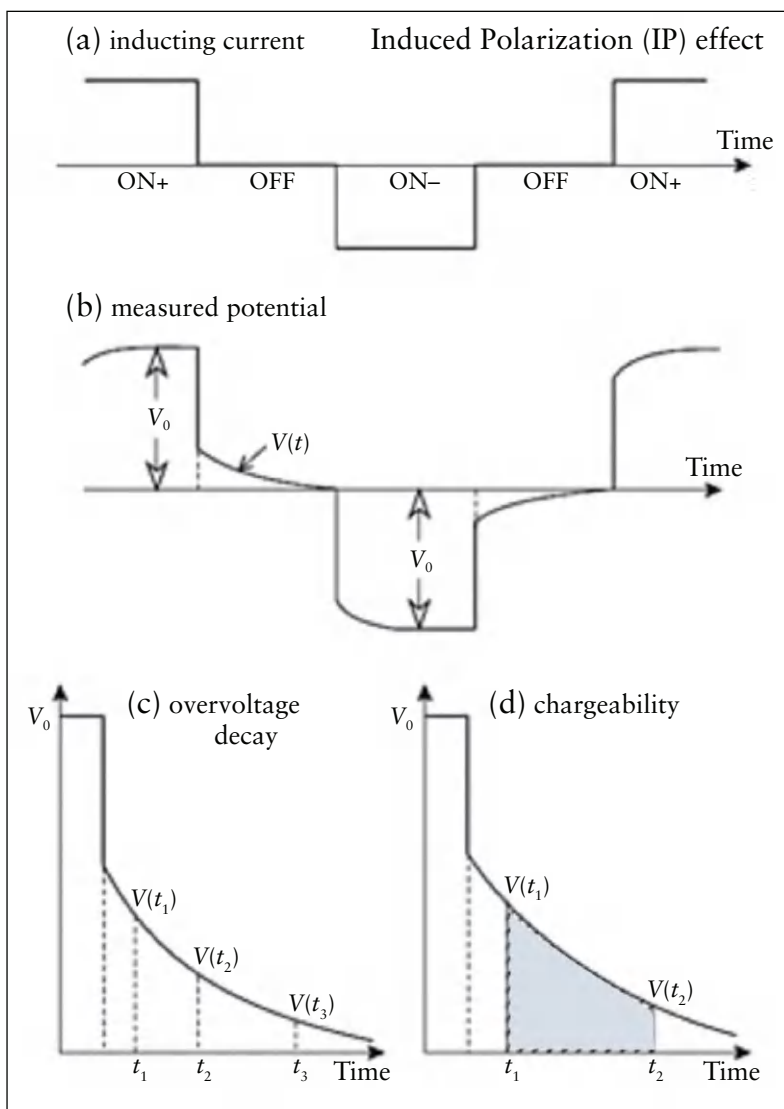


FIGURE 7.1 (a) Illustration of the IP-related decay of potential after interruption of the primary current. (b) Effect of the IP decay time on the potential waveform for a square-wave input current.

Source: Kearey et al. (2002)

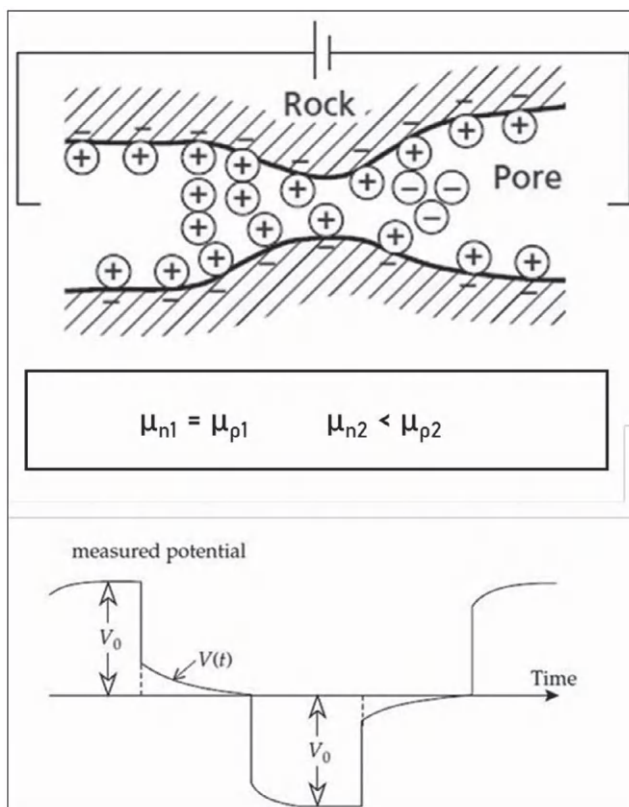


FIGURE 7.2 Membrane polarization associated with constriction between mineral grains.

Source: Kearey et al. (2002).

Membrane polarization is a feature of electrolytic conduction. It arises from the differences in the ability of ions in pore fluids to migrate through a porous rock (Figure 7.2). The minerals in a rock generally have a negative surface charge and thus attract positive ions in the pore fluid. They accumulate on the grain surface and extend into the adjacent pores, partially blocking them. When an external voltage is applied, positive ions can pass through the ‘cloud’ of positive charge, but negative ions accumulate, unless the pore size is large enough to allow them to bypass the blockage. The effect is like a membrane, which selectively allows the passage of one type of ion. It causes temporary accumulations of negative ions, giving a polarized ionic distribution in the rock. The effect is most pronounced in rocks that contain clay minerals, firstly, because the grain and pore sizes are small and, secondly, because clay grains are relatively strongly charged and absorb ions on their surfaces. The ionic build-up takes a short time after the voltage is switched on; when the current is switched off, the ions drift back to their original positions.

Electrode polarization is a similar effect that occurs when ore minerals are present. The metallic grains conduct charge by electronic conduction, while electrolytic conduction takes place around them (Figure 7.3). However, the flow of electrons through the metal is much faster than the flow of ions in the electrolyte, so opposite charges accumulate on facing surfaces of a metallic grain that blocks the path of ionic flow through the pore fluid. An *overvoltage* builds up for some time after the external current is switched on. The size of the effect is commensurate with the metallic concentration. After the current is switched off, the accumulated ions disperse and the overvoltage decays slowly.

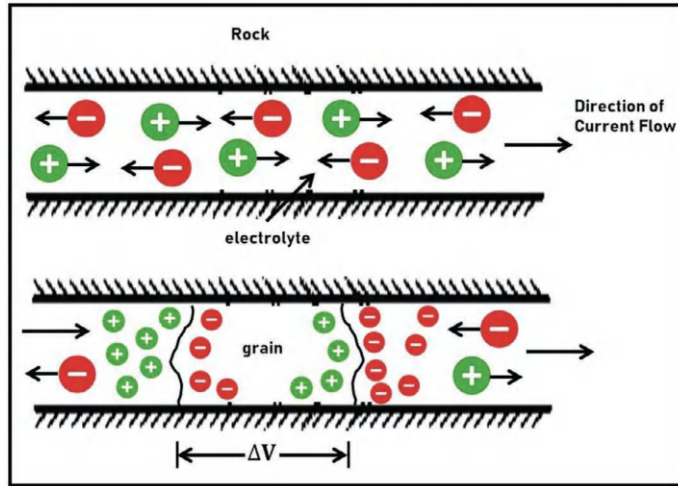


FIGURE 7.3 Electrode polarization IP effect.

Source: Kearey et al. (2002).

The two effects responsible for induced polarization are indistinguishable at measurement level. The field method for an induced polarization (IP) survey is most often based on the double-dipole array. The current electrodes form a transmitter pair, while the potential electrodes form a receiver pair. The steady-state voltage V_0 is recorded and compared with the amplitude of the decaying residual voltage $V(t)$ at time t after the current is interrupted (Figure 7.1c). The ratio $\frac{V_t}{V_0}$ is expressed as a percentage (where $V(t) \ll V_0$, which is expressed as mV/V), which decays during the 0.1–10 s between switching the current on and off. If the decay curve is sampled at many points, its shape and the area under the curve may be obtained (Figure 7.1d). The area under the decay curve, expressed as a fraction of the steady-state voltage, is called the chargeability M , defined as

$$M = \frac{1}{V_0} \int_{t_1}^{t_2} V(t) dt \quad (7.1)$$

M has the dimensions of time and is expressed in seconds or milliseconds. It is the most commonly used parameter in IP studies. The induced polarization determines the length of the potential decay time. If it is shorter than the time when the inducing current is off, successive half-cycles of the potential will not interfere. However, if a disseminated conductor is present, the decay time increases, causing overlap and distortion of the half-cycles. The higher the signal frequency, the more pronounced the effect. It increases the ratio $\frac{V(t)}{V_0}$, giving the impression of a better conductor than is really present (i.e. the apparent resistivity decreases with increasing frequency). Clearly, IP and resistivity surveys with alternating current are also influenced. The frequency dependence of the IP effect is exploited by measuring the apparent resistivity at two low frequencies. Let these be f and F ($F > f$). Commonly, $f \approx 0.05$ –5 Hz and $F \approx 1$ –10 Hz. Then, $\rho_f > \rho_F$, and we can define a *frequency effect* as

$$FE = \frac{\rho_f - \rho_F}{\rho_F} \quad (7.2)$$

The ratio FE is often multiplied by 100 to express it as a percentage (PFE). If no IP effect is present, the resistivity will be the same at both frequencies. The larger the value of FE or PFE , the greater the induced polarization in the ground. At frequencies above 10 Hz, mutual inductance effects between the cables of the primary and detection circuits can produce troublesome potentials, which must be avoided by the field procedure (such as restricting F) or minimized analytically.

The presence of metallic conductors is expressed by a similar parameter to FE , the metallic factor (MF). This is proportional to the difference in conductivities at the two measurement frequencies (Marshall and Madden, 1959).

$$MF = A (\sigma_F - \sigma_f) = A (\rho_f^{-1} - \rho_F^{-1}) = A \left(\frac{\rho_f - \rho_F}{\rho_f \rho_F} \right) \quad (7.3)$$

The constant A is equal to $2\pi \times 10^5$; the units of MF are those of conductivity (i.e. $\Omega^{-1} \text{m}^{-1}$ or S m^{-1}).

Relation between time domain (TD) and frequency domain (FD) IP measurement, as suggested by Siegen (in Wait (1959a, b, c)):

$$MF = \frac{PFE}{100 + PFE} \quad (7.4)$$

7.3 PHASE DOMAIN (MEASURE PHASE DELAYS BETWEEN CURRENT AND VOLTAGE)

If the transmitting system sending the current and the receiving system measuring the voltages are time-linked, the phase difference between the transmitted and received signals is measured. This difference, measured in terms of time or phase angle, gives the polarization of the intervening Earth. As phase angles are very small, their unit of measurement is milliradians, and they are the same as chargeability as well as PFE in defining the IP effect. These angle measurements are usually made by determining the in-phase and out-phase components and then finding the tangent of the defined angle.

Conventional IP measurements, even when a large range of frequencies is utilized, measure only the absolute volume without regard to in-phase and out-phase or vector components. But in order to measure all effects truly and to be able to separate the response from one type to another type of polarizable material, measurements of the in-phase and out-phase IP components must be made over a spectrum of wide range of frequencies. Such measurements are known as complex resistivity measurements. The complex resistivity spectrum of material is shown to be governed by the nature of the medium and may be used as a means for mineral discrimination, and the technique is termed *spectral induced polarization* (SIP), using AC current and measure phases and amplitudes at frequencies 30 to 4×10^3 Hz (Pelton et al., 1978). Measuring *complex impedance*:

$$Z(\omega) = \frac{U(\omega)}{I(\omega)} K \quad (7.5)$$

where K is the geometric factor of the array.

Complex impedance measurements of materials have been made in studies of dielectric phenomenon at least since 1941 (Cole and Cole, 1941). With the advent of microprocessor technology in the field of electronics, field measurements of multifrequency phase and amplitude IP response are made possible by present-day instrument.

The SIP method is based on using the Cole–Cole dispersions (i.e. the spectra of a Cole–Cole model) to fit the field complex resistivity data. The Cole–Cole model, as it is termed, first proposed by Cole and Cole (1941) to predict complex dielectric behaviour, was used by Pelton et al. (1978) to predict complex resistivity behaviour. The *Cole–Cole* model for complex resistivity (Figure 7.4):

$$Z(\omega) = \rho_0 \left[1 - M \left(1 - \frac{1}{(1 + i\omega\tau)^c} \right) \right] = |Z| e^{i\varphi} \quad (7.6)$$

where ρ_0 is the DC resistivity, M is the chargeability, and τ is the relaxation time.

For varying frequencies, complex resistivity describes a semicircle in the $(\text{Re } Z, \text{Im } Z)$ plane (Cole–Cole relaxation spectra) (Figure 7.5):

$$Z(\omega) = \rho_0 \left[1 - M \left(1 - \frac{1}{(1 + i\omega\tau)^c} \right) \right] = R(Z) + i I(Z) \quad (7.7)$$

The critical frequency at which the maximum phase shift is measured is indicative of τ :

$$F_c = \frac{1}{2\pi (1 - M)^{\frac{1}{2c}}} \quad (7.8)$$

which is independent of resistivity. The phenomenon of polarization in both the domains is the same, and chargeability as well as PFE reflect the electric polarizability of the medium. More information on spectral IP can be found in Kemna (2000), Kemna et al. (2012), and Binley (2015).

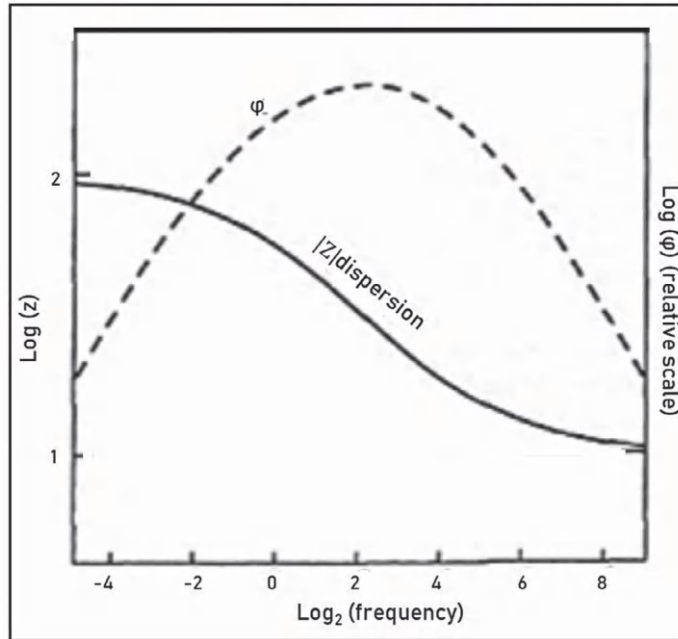


FIGURE 7.4 Cole–Cole model for complex resistivity.

Source: Pelton et al. (1978).

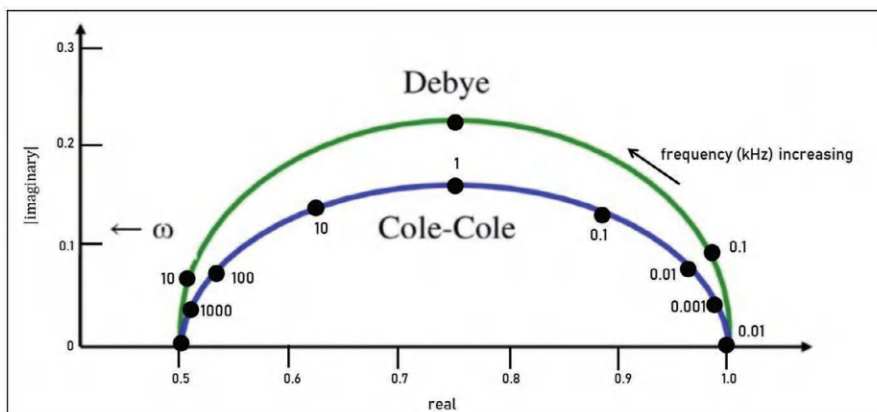


FIGURE 7.5 Cole–Cole relaxation spectra.

Source: Pelton et al. (1978).

7.4 THE MAGNETIC INDUCED POLARIZATION (MIP) METHOD

In a normal IP survey (Seigel and Howland-Rose, 1990), as the secondary voltages are screened off by extreme conditions of overburden conductivity, it is reasoned that if the magnetic effects of the primary and secondary current flows in the ground can be measured, the difficulty of conductive overburden will be overcome, which results in the development of the MIP method (Seigel, 1974). A sensitive vector magnetometer measures the magnetic field caused by the primary current flow as well as the secondary currents both inside and outside the body (Chen et al., 2002). As the magnetic field is not at all hampered by the overburden conductivity, the problem of screening off the signal by surface conducting layers is solved by this method. The measurements are made either in the time or frequency domain, though the latter is better suited as it offers better signal-to-noise ratio. The field procedure is slightly different from that of electrical-induced polarization (EIP).

The ground is energized by two widely placed electrodes parallel to the strike direction, and the primary and secondary magnetic fields are measured from the analogue output of a sensitive vector magnetometer by the normal IP receiver. The magnetometer is placed at successive stations on traverse lines perpendicular to the strike as usual. The data is plotted as profiles, the main parameters being primary magnetic field (γ) and chargeability (milligamma/gamma). The primary magnetic field represents the conductivity variation along the traverse. The MIP methods might form a supplement to electric induced polarization (EIP) method in specific areas of high overburden conductivity.

It is well-known that one of the benefits of MIP measurement over EIP is that MIP works best in conductive areas, where it is normally difficult to measure voltages with EIP. This scenario can be envisioned in the schematic shown in Chen et al. (2002) and Seigel and Howland-Rose (1983), which was studied with a scale model in Seigel and Howland-Rose (1983). Since the overburden is highly conductive, it is expected that the applied current on the surface can hardly penetrate this layer and energize the target below. Both MIP and EIP might have difficulties in obtaining good responses. One remedy is to locate the source electrodes down below the overburden so that the galvanic current will be channelled directly into the target. Due to the masking effect of the overburden, EIP still has a problem to pick up the weak electrical signal. However, if we measure the magnetic signal instead, then it seems that there is no obstacle at all.

7.5 CHARGEABILITY OF VARIOUS MATERIALS

The following are the chargeability (Table 7.1) and metal factor (Table 7.2) of the various materials.

TABLE 7.1
Chargeability of Various Materials

Material	Chargeability (ms)
Groundwater	0
Alluvium	1–4
Gravels	3–9
Precambrian volcanics	8–20
Precambrian gneisses	6–30
Schists	5–20
Sandstones	3–12
Argillites	3–10
Quartzites	5–12

TABLE 7.2
Metal Factor of Various Rocks and Minerals

Material	Metal Factor (mhos/cm)
Massive sulphides	10,000
Fracture-filling sulphides	1,000–10,000
Massive magnetite	3–3,000
Porphyry copper	30–1,500
Disseminated sulphides	100–1,000
Shale sulphides	3–300
Sandstone, 1 to 2% sulphides	2–200
Finely disseminated sulphides	10–100
Tuffs	1–100
Graphitic sandstone and limestone	4–60
Gravels	0–200
Alluvium	0–200
Precambrian gneisses	10–100
Granites, monzonites, diorites	0–60
Various volcanics	0–80
Schists	10–60
Basic rock (barren)	1–10
Granites (barren)	1
Groundwater	0

7.6 IP INSTRUMENTS

- The TD IP equipment consists generally of a petrol-driven motor generator normally of 400 Hz frequency at 110 or 210 V.
- Transmitter and receiver unit, cables, current electrode, and porous pots.
- High-power current generates 10 KW and operates in the range of 1,000–10,000 V, with a maximum current output of 15–20 A.
- Low-power 3 KW operates in the voltage range 200–1,500 V, with a maximum current of 3 A.
- The transmitter provides square-wave pulse with equal on–off times, with a choice of 1, 2, 4, or 8 seconds (Figure 7.6).

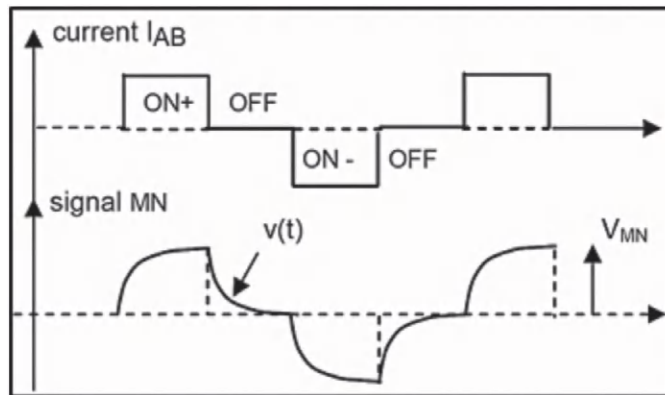


FIGURE 7.6 Square-wave pulse with equal on–off times.

Source: iris-instruments.com (VIP 10 000 Tx and ELREC Pro Rx), 1 Avenue Buffon, BP 6007, 45060 Orléans Cedex 2, France.

- The receiver consists of automatic self-potential bucking device with a digital display for *SP*, primary voltage, as well as chargeability.

With the ELREC Pro, ten dipoles are measured at the same time (*dI* to *dI0*), which decreases the duration of the survey.

7.7 IP FIELD SURVEY, PROCESSING, AND INTERPRETATION OF DATA

Field procedure and equipment are the same as the resistivity exploration. The fieldwork is a slow process compared to that of magnetic and *SP*. IP surveys are generally followed by *SP* and magnetic. Electrode array may be the same as resistivity survey, namely, Wenner, Schlumberger, gradient, pole–dipole, and dipole–dipole. Some commonly used electrode array configurations are included in Figure 7.7.

The selection of a particular electrode configuration for a particular area depends upon the requirement of the resolution, the overburden conditions, the attitude of the ore body, etc. As high overburden conductivity gives pronounced inductive coupling effects, an array with minimum EM coupling may have to be used in such areas. A short summary of each mentioned configuration in Figure 7.7 will be realized regarding the three parameters (sensitivity, depth of investigation [DOI], signal-to-noise ratio [S/N]) in order to explain and understand the choice for this study, as well as to get a grasp on what is at stake when choosing a particular array.

1. Wenner Array

- Comment: Robust and traditional array.
- Sensitivity: This array is more sensitive than vertical variations in the subsurface. As can be seen on Frederic and Robin (academic year 2016–2017; modified after Loke, 1999), the contours are aligned horizontally. Obviously, it is weaker in detecting horizontal gradients.
- DOI: The median depth of investigation is about $\frac{a}{2}$, where a is the electrode spacing. It thus has a moderate DOI compared to other configurations.
- S/N: This ratio is inversely proportional to the geometric factor K , and the Wenner array has one of the smallest values among the different arrays ($2\pi a$), conferring it the highest signal strength.

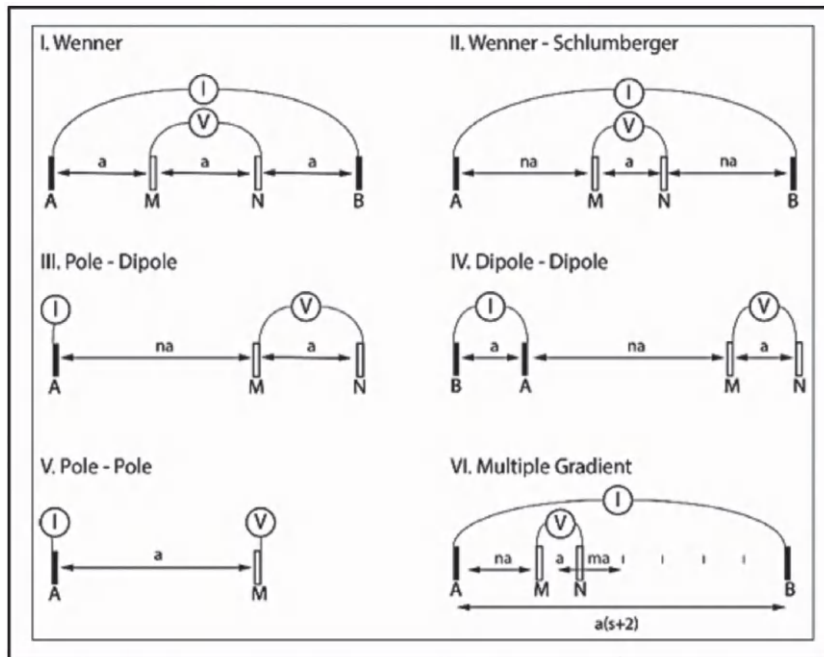


FIGURE 7.7 Some commonly used electrode configuration arrays: a is the distance between electrodes, and the factors m , n , and s are scalars.

Source: Combined from Loke (1999) and Dahlin and Zhou (2006).

2. Wenner–Schlumberger Array

- Comment: Hybrid between the Wenner and Schlumberger array (not described here).
- Sensitivity: Compared to the Wenner array sensitivity figure, there is a slim vertical curvature beneath the middle of the array, and lower sensitivity values in the zones between M and A and, symmetrically, B and N (Frederic and Robin, academic year 2016–2017; modified after Loke, 1999). According to the n factor, the configuration will be moderately sensitive to horizontal (low n) and vertical (higher n) variations.
- DOI: The median DOI is about 10% deeper than for the Wenner array for an equal distance between electrodes A and B .
- S/N: The S/N ratio is smaller than for the Wenner array but better than the dipole–dipole array's.

3. Pole–Dipole Array

- Comment: Appealing alternate option for IP investigations compared to the dipole–dipole array (increased signal strength) and to the Wenner/Wenner–Schlumberger arrays (lower electromagnetic coupling).
- Sensitivity: The best sensitivity values lie below the potential electrodes M and N . For an increasing n factor, the elevated sensitivity zone below the potential electrodes becomes more and more aligned with the vertical axis that it becomes adequate for detecting vertical gradients (Frederic and Robin, academic year 2016–2017; modified after Loke, 2015).
- DOI: It goes slightly deeper than the dipole–dipole array, and it is possible to increase it by augmenting the parameter a .
- S/N: It has a substantially bigger S/N ratio than that of the dipole–dipole array and is not as susceptible to pick telluric noise as the pole–pole array.

4. Dipole–Dipole Array

- Comment: Extensively used in resistivity and IP investigations on account of the small electromagnetic coupling between the circuits.
- Sensitivity: The zones containing the best sensitivities are located beneath the two electrodes couples (Frederic and Robin, academic year 2016–2017; modified after Loke, 1999). The two lodes formed by these high sensitivities converge towards the middle of the array. The contour patterns tend to come closer to vertically as the factor n increases. It results in a highly adequate configuration for picking horizontal resistivity gradients in the subsurface. As a trade-off, its capacity to detect vertical gradient is weaker.
- DOI: Its penetration depth depends on the n factor. It usually has a lower DOI compared to the Wenner array.
- S/N: A particular weakness of this configuration is the terrible signal strength for large n factor values, since the potential and current couples are then so far from each other.

5. Pole–Pole Array

- Comment: Prominent in specific applications, such as archaeological investigations, where relatively short electrode spacings are used.
- Sensitivity: The large spacings between the contours of the 2-D sensitivity plot indicate that this method has a poor resolution (Frederic and Robin, academic year 2016–2017; modified after Loke, 2015).
- DOI: It has the deepest depth of investigation of all the arrays.
- S/N: The second current and potential electrodes B and N are actually located at a great distance from A and M . Because of this considerable separation, this configuration is susceptible to collect generous amounts of telluric noise, altering the data quality.

6. Multiple-Gradient Array

- Comment: Designed in 2006 by Dahlin and Zhou. Several series of voltage gradient measures are realized with the current electrodes at diverse points. It can be envisioned as englobing the attributes of the pole–dipole and the Wenner–Schlumberger.
- Sensitivity: For fixed A and B potential electrodes and moving potential electrodes M and N , the sensitivity pattern will evolve from a Wenner–Schlumberger array type for M , N far from A and B , to a pole–dipole array type when the potential couple comes close to either A or B . The figure in Frederic and Robin (academic year 2016–2017; modified after Loke, 2015) is an intermediate between those two situations, when M and N approach B .
- DOI: It is comparable to the pole–dipole and Wenner–Schlumberger arrays.
- S/N: The S/N ratio is more favourable than in the case of the dipole–dipole or pole–dipole arrays (Dahlin and Zhou, 2006).

The following table (Table 7.3) shows a comparison of the commonly used electrode configurations for IP surveys (Coggon, 1973).

Considering the advantages in anomaly resolution, depth of exploration, horizontal location, and freedom from EM coupling effects, the dipole–dipole array gained maximum popularity among field geophysicists despite the accompanying ‘low signal-to-noise ratio’ problem. While the choice of the electrode array is mainly, generally based on practical considerations, the choice of the electrode spacing is more important. An improper choice of electrode interval can result in an apparent anomaly that may be too small to be separated from background variations, or no anomaly at all. In a general way for a shallow-occurring source, the closer the electrode spacing to the width of the body, the greater will be the apparent IP effect. Measurement at shorter spacings

TABLE 7.3

Comparison of Common IP Arrays

Serial No.	Characteristic	Performance		
		Dipole–Dipole	Three-Electrode	Gradient
1	Magnitude of response	Good	Fair	Poor
2	Dip information	Poor	Fair	Good
3	Depth of exploration	Good	Good	Good
4	Horizontal resolution	Fair	Poor	Good
5	Target location	Good	Fair	Good
6	Freedom from overburden irregularities	Fair	Good	Poor
7	Freedom from EM coupling	Good	Fair	Poor
8	Depth estimation	Fair	Fair	Poor
9	Signal-to-noise ratio	Poor	Fair	Good
10	Speed of coverage	Fair	Fair	Good
11	Labour needed	Poor	Fair	Good

can be expected to detect shallow sources. Although it is possible to ‘look to greater depth’ by increasing the electrode interval, it must be remembered that measurements with larger electrode spacings average a large volume of rock into each measurement, and the apparent anomaly from a small, possibly important, source may be missed. Generally, the layout is a 10 m station interval and 50 m traverse interval. For reconnaissance, it may be done with 50 m station interval and 200 m traverse interval.

An IP survey includes both lateral profiling and vertical sounding with the expanding spread method. Using a double–dipole array, the distance between the nearest electrodes of the transmitter and receiver pairs is a multiple (na) of the electrode spacing a in each pair. Measurements are made at several discrete positions as the receiver pair is moved incrementally away from the fixed transmitter pair. The transmitter pair is then moved by one increment along the profile, and the procedure is repeated. The value of ρ_a , FE , or MF obtained in each measurement is plotted below the midpoint of the array at the intersection of two lines inclined at 45° (Figure 7.8a). Information is obtained from increasingly greater depths as the transmitter–receiver array expands (i.e. as n increases). The plotted value is, however, not the real value of the parameter at the indicated depth (recall, for example, that a measurement of apparent resistivity represents an equivalent half-space beneath the array). A two-dimensional picture of the variation of the IP parameter beneath the profile is synthesized by contouring the results (Figure 7.8b). The plot is called a *pseudosection*; it provides a convenient (though artificial) image of the presence of anomalous conductors but does not represent their true lateral or vertical extent. The presence of anomalous regions may be investigated further by exploratory drilling.

Resistivity anomalies depend on the presence of continuous conductors, such as groundwater or massive ore bodies. If mineralization is disseminated through a rock, it may not cause a significant resistivity anomaly. The good response of the IP method for disseminated concentrations of conducting ore minerals led to its development in base metal exploration, where large low-grade ore bodies may be commercially important. However, the IP effect also depends on the porosity and saturation of the rock. As a result, it can also be used in the search for groundwater and in other environmental applications. The qualitative interpretation of the IP field results (profiles or pseudosections) yields a large amount of required information. Location, lateral extent, and depth of indications can be judged from the characteristics of the anomalies, such as sharpness, symmetry, and amplitude above the background. For the quantitative approach, the technique of interpretation in recent times is by computer modelling. In general, there are two main computer

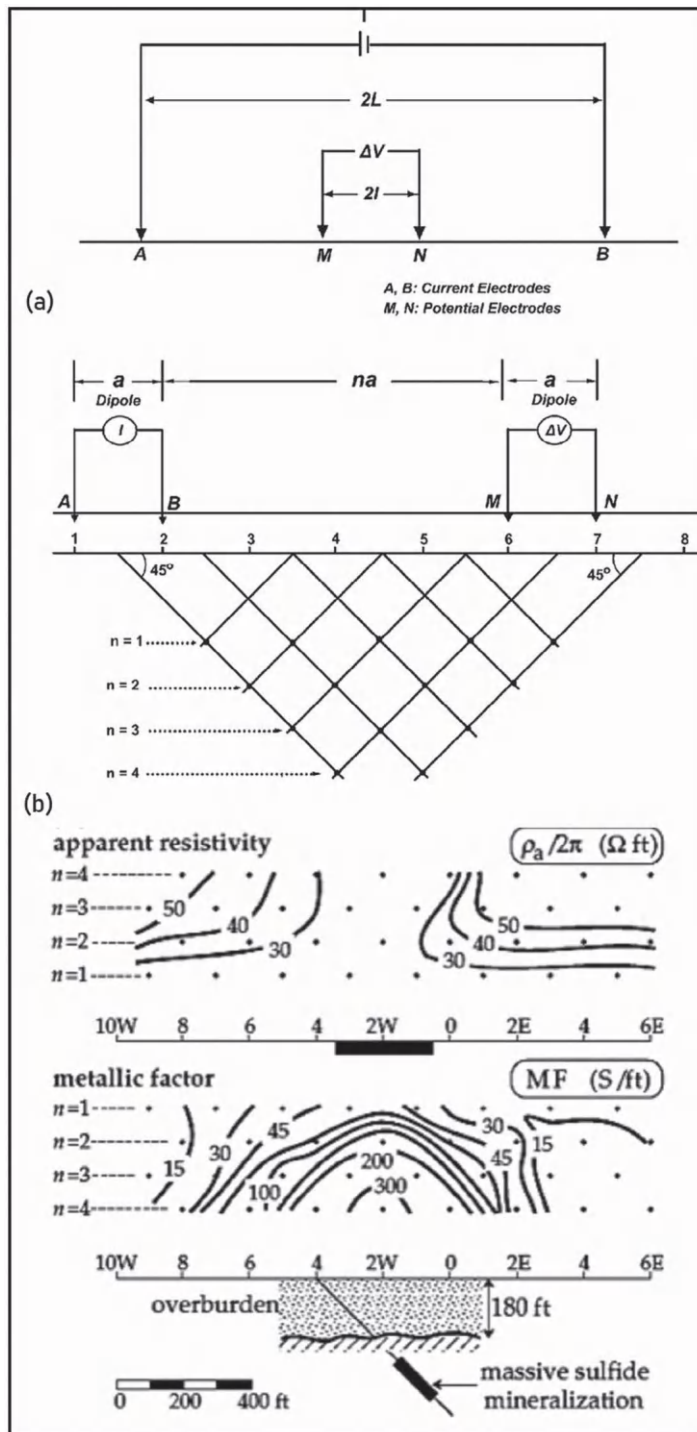


FIGURE 7.8 (a) Construction of a pseudosection for a dipole-dipole IP survey: the measured parameter is plotted at the intersection of 45° lines extending from the midpoints of the transmitter and receiver pairs. (b) Pseudosections of apparent resistivity and metallic factor for an IP survey over a sulphide ore body.

Source: Telford et al. (1990).

approaches for modelling, which can be called the forward solution and the inverse solution. In the former case, the interpreter normally interacts with the computer and selects a model of the IP response which best fits with the field response, and the latter is a direct interpretation method, where the observed data is analyzed to obtain a direct solution of the subsurface. Several modelling techniques are discussed in published literature (Coggon, 1971; Aiken and Hastings, 1973). However, in spite of all the sophistications and advancements by computer aids, the interpreter must always be guided by geologically reasonable boundary conditions, intuition, and of course, experience.

IP soundings are most commonly made using the Schlumberger array, pole-dipole array, or Wenner array, and usually in the time domain. The apparent chargeability η_a versus the electrode spacing a is plotted on logarithmic coordinates. The IP sounding curve is an interpreted curve matching procedures, either graphically, using sets of IP sounding master curves, or by computer. At present, only a few two-layer master curves (for the Wenner array) have been published in the United States (Seigel, 1959; Frische and von Buttlar, 1957). Three- and four-layer curves have been published in the Soviet Union. An IP sounding curve can be of significant value in complementing a resistivity sounding curve. For example, the resistivity and IP sounding curves for the following four-layer geoelectric section are shown in Figure 7.9.

It is obvious that layer 3 cannot be distinguished on the four-layer resistivity curve (which resembles a two- or three-layer curve). But layer 3 is characterized by a different chargeability from the surrounding layers, and its presence is indicated clearly by the IP sounding curve.

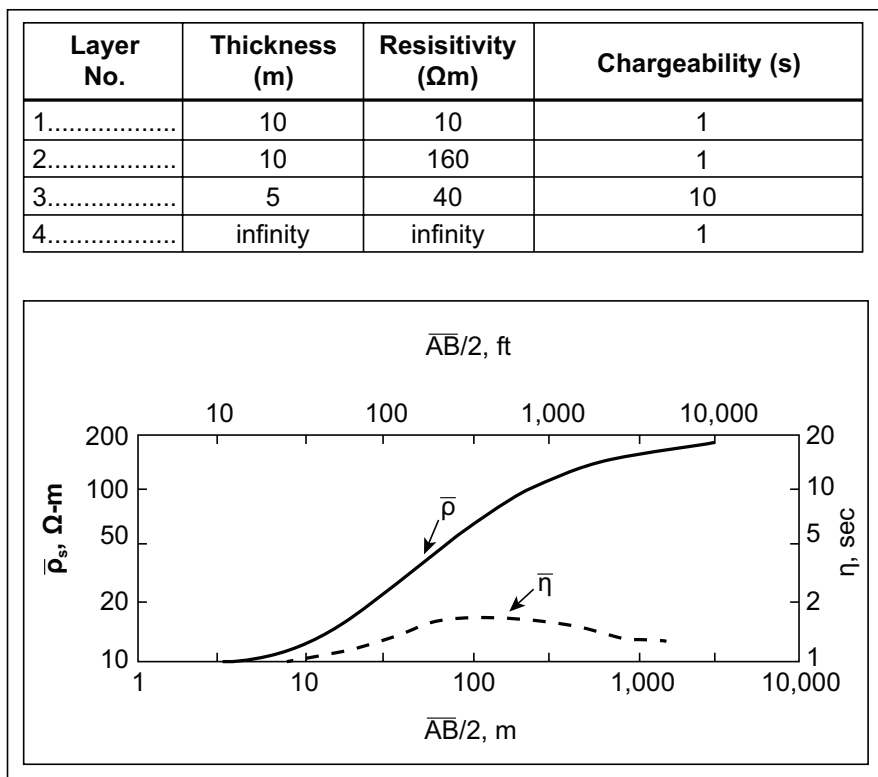


FIGURE 7.9 Apparent resistivity and apparent chargeability (IP) sounding curves for a four-layer model.

Source: Zohdy (1974a, 1974b).

7.8 CROSS-BOREHOLE IMAGING

Cross-borehole imaging, or tomography, employs similar techniques as those used in soundings and profiling, resulting in an electrical cross section of the resistivity or IP property distribution. The technique has been used for many different applications, including permeable reactive barrier (PRB) integrity imaging, hydrocarbon contaminant mapping, groundwater infiltration time series imaging, tracer studies, permeability and transmissivity estimation, etc. The figure in Martinho (2013; modified after Slater et al., 2007; modified after Kemna et al., 2000) presents an example of resistivity and IP tomography data taken from a hydrocarbon-contaminated (kerosene) site. The soil column shows a fine-grained layer at depth, imaged by the resistivity data. The IP data suggests that this fine-grained layer appears to retard the flow of hydrocarbon within the subsurface.

7.9 LIMITATIONS OF IP

The limitations of the normal IP surveys are obvious. In case of weak responses, it is generally not possible to separate the anomaly due to a water-filled shear zone from that of a mineral. Similarly, it is not possible to separate the chargeability or PFE anomalies caused by uneconomic metallic conductors from those of the economic ones. Another drawback, as with the other electrical methods, is that the response from a causative source under a highly conducting overburden will generally be masked by the latter and left undetected by IP surveys. In an attempt to solve these problems, considerable research and development effort has been made, which resulted in major advancements in the IP methodology, a brief account of which is given in what follows.

7.10 MEASUREMENT ERRORS

Electric investigations can be affected by several sources of noise and errors. Here is a non-exhaustive list englobing the main sources of noise. For more details, see Sumner (1976).

- *Telluric noise.* Low-amplitude and gently evolving currents flow continuously through the Earth (Sumner, 1976). Their direction may vary but is typically parallel to the ground surface. However, their effects are easily suppressed when using alternating current, which cancels their effects. Indeed, at each direction change of current, the telluric current alternately increments or decrements the measured potential difference by identical magnitude; summing the output over various cycles removes the detrimental effect (Frederic and Robin, academic year 2016–2017; modified after Kearey et al., 2002).
- *Man-made electrical noise.* Luckily, hardly any man-made electrical instruments operate in the IP frequency band of interest (Sumner, 1976). Utilizing filters to diminish undesirable noise is an appropriate way of augmenting the signal-to-noise ratio of the IP receiver. For example, low-pass filtering may be used to reject power line frequencies and radio station interference, at the threshold of 10 Hz, since there are usually no IP frequencies of interest above that value.
- *Sferics.* They are electrical phenomena originating from the atmosphere. They essentially consist of electromagnetic pulses due to thunderbolts. Their frequency band is normally higher than that of the IP but can be troublesome. If the data become affected by this event, a thunderstorm could be too close (Sumner, 1976). This somewhat-anecdotal source of noise is only mentioned here because thunderstorms forced the crew to flee the field at several occasions during the investigations of this study, although detrimental aspects of sferics have not been observed on the data, or its aspects could not have been differentiated from other sources.

- *Self-potential noise.* Self-potential voltages result from electrochemical reactions within the subsurface, in the neighbourhood of the profile. They typically have very low amplitudes and are incoherent in their frequencies (Sumner, 1976).
- *Contact resistance.* The contact resistance between the ground and the electrodes can lead to bad readings. However, the measure instrument computes those resistance beforehand, allowing the user to improve this issue, for example, by hammering the electrodes deeper and/or by adding some electrolytic fluid in a relatively small zone around the concerned electrodes (Binley, 2015).
- *Measured potentials magnitude.* During data acquisition, it is important to check if the measured voltages are high enough, that is, not too close to the instrument resolution. This can occur in practice when the input voltage is too weak and/or when associated with big geometric factors. This can be problematic in IP investigations where a secondary, decaying voltage is measured, whose magnitude is largely below the primary voltage. This is why the input current needs to be more intense and applied longer (a few seconds) than a DC survey (Binley, 2015).
- *Electrode polarization.* This phenomenon can lead to spurious readings in IP investigations when an intense current is injected for a prolonged time. An ingenious attempt to reduce electrode polarization is to regularly switch the polarity of the current electrodes (Binley, 2015).
- *Coupling of electrical circuits.* This phenomenon can cause errors during IP investigations; it refers to the resistive and capacitive coupling between an IP transmitter and receiver (Sumner, 1976). The capacitive coupling is a voltage variation owing to:
 1. Electrical leakage from transmitter or receiver electrodes to receiver or transmitter wires. This effect can change the charge distributions stirring capacitance.
 2. Displacement currents from wire to wire or wire to the ground. To lessen capacitive coupling, the receiver and transmitter should be at a sufficiently large distance between each other (Sumner, 1976). This is why the dipole–dipole configuration has originally been favoured, due to the large distance between the transmitter and the receiver (Binley, 2015).
- *Cultural coupling effect.* This detrimental effect appears when a man-made metallic (conductive) object is in the neighbourhood of the electrode layout. It can lead into spurious IP anomalies readings, mainly due to leakage currents between the transmitter and the receiver circuits. If detected, some ways to overcome the effects are to somehow insulate the object, by using lower frequencies, or simply removing it if possible (Sumner, 1976).
- *Electromagnetic coupling.* In some circumstances, the IP transmitter and receiver act like the primary and secondary coil windings of a usual electrical transformer—the primary circuit induces a current in the secondary circuit, phenomenon aggravating with higher frequencies (Sumner, 1976). This electromagnetic induction provokes deceptive IP-similar effects. The effect is more pronounced in conductive environments and long-array geometries chosen to investigate deeper bodies but is more.

7.11 APPLICATIONS AND CASE HISTORIES OF IP SURVEY

The induced electrical polarization method is widely used in the exploration for ore bodies, principally of disseminated sulphides. The use of IP in geotechnical and engineering applications has been limited, and it has been used mainly for groundwater exploration. Groundwater IP studies generally have been made with time domain IP. Environmental examples of the successful use of IP include saline intrusion mapping (Seara and Granda, 1987), the detection of clay units (Iliceto et al., 1982; Vinegar and Waxman, 1984), the detection of both inorganic

contaminants (Cahyna et al., 1990; Ruhlow et al., 1999) and organic contaminants (Olhoeft, 1984, 1985, 1992; Vanhala et al., 1992), and permeability estimation (Knoll et al., 1994; Borner et al., 1996; Sturrock et al., 1999). Although IP methods are being used more frequently in environmental investigations, they are not as widely used as other electrical methods (e.g. DC resistivity, electromagnetic [EM] methods, and ground-penetrating radar), and their full potential has yet to be realized.

7.11.1 MAPPING SOIL AND GROUNDWATER CONTAMINATION: IDENTIFICATION OF A CONTAMINATION WITH CYANIDE COMPLEXES (SLAGS FROM PLATING WORKS), CZECHOSLOVAKIA

Cahyna et al. (1990) show a valuable IP example used to determine the slag-type material containing cyanide complexes that have contaminated groundwater in Czechoslovakia. Figure 7.10 shows the contours of η_a (percent) obtained from a 10 m grid of profiles (Cahyna et al., 1990).

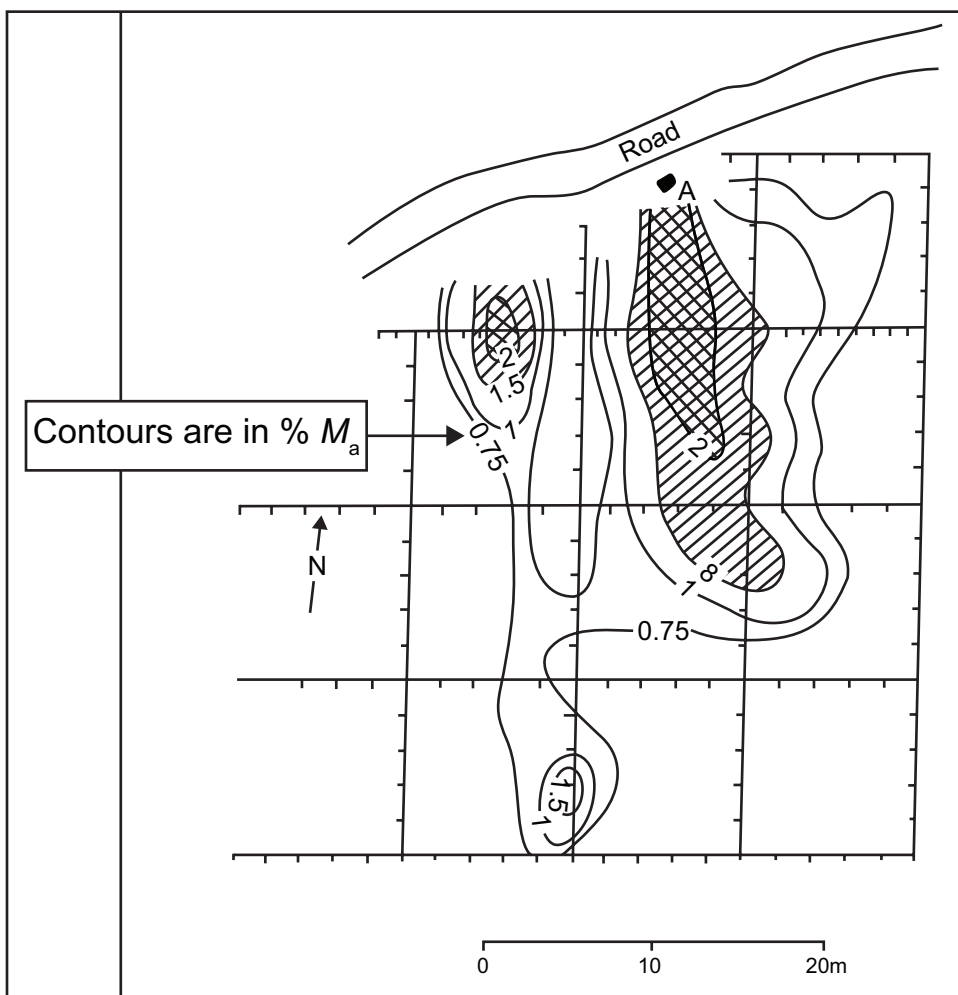


FIGURE 7.10 Network of SRP-IP profiles with contours of IP vs η_a (%) and extent of contaminant interpreted on the basis of the geophysical survey.

The largest IP anomaly ($\eta_a = 2.44\%$) directly adjoined the area of the outcrop of the contaminant (labelled A). The hatched region exhibits polarizability over 1.5% and probably represents the maximum concentration of the contaminant. The region exhibiting polarizability of less than 0.75% was interpreted as ground free of any slag-type contaminant. The resistivity survey failed to detect the contamination, whereas IP chargeability identified both the known and unknown slag deposits.

7.11.2 GROUNDWATER EXPLORATION WORK NEAR THE SAUK-SOO RIVER IN CRIMEA AND IN THE KALININO REGION OF ARMENIA

The majority of published case histories using IP surveys have been for mining exploration, but those treating groundwater exploration are growing: Vacquier et al. (1957), Kuzmina and Ogilvy (1965), Bodmer et al. (1968), and Sternberg and Oehler (1990). Kuzmina and Ogilvy reported on work done near the Sauk-Soo River in Crimea and in the Kalinino region of Armenia. In Crimea, the IP work consisted essentially of IP sounding (time domain) using the Wenner array. The alluvial deposits in the studied area were poorly differentiated by their resistivities, but three horizons were clearly distinguished by their polarizability (Zohdy, 1974b; modified after Kuzmino and Ogilvi, 1965). The section consisted of a top layer of weak polarizability ($h_1 = 2\text{--}4\text{ m}$; $\eta_1 = 0.8\text{--}1.5\%$), which represents a dry loamy layer; a second layer of strong polarizability ($h_2 = 8\text{--}20\text{ m}$; $\eta_2 = 3\text{--}5\%$), which represented a clayey sand layer saturated with fresh water; and a third layer of weak polarizability (h_3 , very thick; $\eta_3 = 1\%$), which represents impervious siltstones. The survey in this area demonstrates that the IP work provided more complete information about the groundwater occurrence than did the resistivity soundings alone.

7.11.3 IP INTERPRETATION IN ENVIRONMENTAL INVESTIGATIONS, CRESCENT BEACH STATE PARK, NEAR PORTLAND, MAINE

A time domain IP survey was used to investigate the structure of the freshwater–saltwater interface at Crescent Beach State Park, near Portland, Maine (Slater and Sandberg, 2000). The figure in Slater and Lesmes (2002) shows apparent resistivity ρ_a , apparent chargeability M_a , and apparent normalized chargeability MN_a pseudosections for a dipole–dipole survey line extending from 20 m above the high tide mark to 60 m below the high tide mark. The inverted models are shown in Slater and Lesmes (2002) and Slater and Sandberg (2000). The freshwater–saltwater interface is revealed as the sharp resistivity transition immediately up-beach from the high tide mark. Up-beach of 142 m, the chargeability increases up to a maximum of 9.6 mV/V in the inversion (Slater and Lesmes, 2002). However, both the MN_a pseudosection and the inverted MN model are uniform across this interface.

7.11.4 DETECTION OF METAL PIPES AND CABLES IN BAIMA, CHINA

Luo and Zhang (1998) show model experiments and analytical results that show that, in certain circumstances, a buried metal pipe or armoured cable can introduce anomalies in IP (and apparent resistivity) with large amplitude and wide range. These results are important for two reasons. The first is that such man-made features have the potential to cause ‘noise’ or errors in electrical surveys. The general rule of thumb when planning a survey is to orient soundings and profiles as nearly perpendicular to any known buried pipes or cables as the field conditions allow. The figure in Wightman et al. (2003; EPA environmental geophysics; Luo and Zhang, 1998) shows the results of an IP survey using the gradient array in Baima, China. An η_a

anomaly of 10% to 3% with a width of more than 200 m was obtained near the road (the dot-and-dash lines in the figure). This anomaly can be traced for 4 km along the road. The trend of the anomalies is basically consistent with the road and independent of the stratum strike or structural direction within the prospect area. Apparently, it results from a buried communication cable along the highway rather than geological features. The apparent resistivity profiles (the dot-and-double-dash lines on the figure) also appear to correlate with the cable, but with much less consistency or amplitude.

7.11.5 LABORATORY CASE HISTORY

Ogilvy and Kuzmina (1972) carried out laboratory experiments with mixtures of quartz sand and various clay fractions for each of several clay minerals. Maximum polarizations were observed in the range 3–8 wt% of clay (Figure 7.11a).

At greater clay contents, the induced polarization effect gradually decreased with an increase in clay content. The lowest IP values related to pure quartz sand and pure clays. Figure 7.11b shows the variation of chargeability in freshwater and saline water lenses. The freshwater lens is well detected by a high chargeability anomaly.

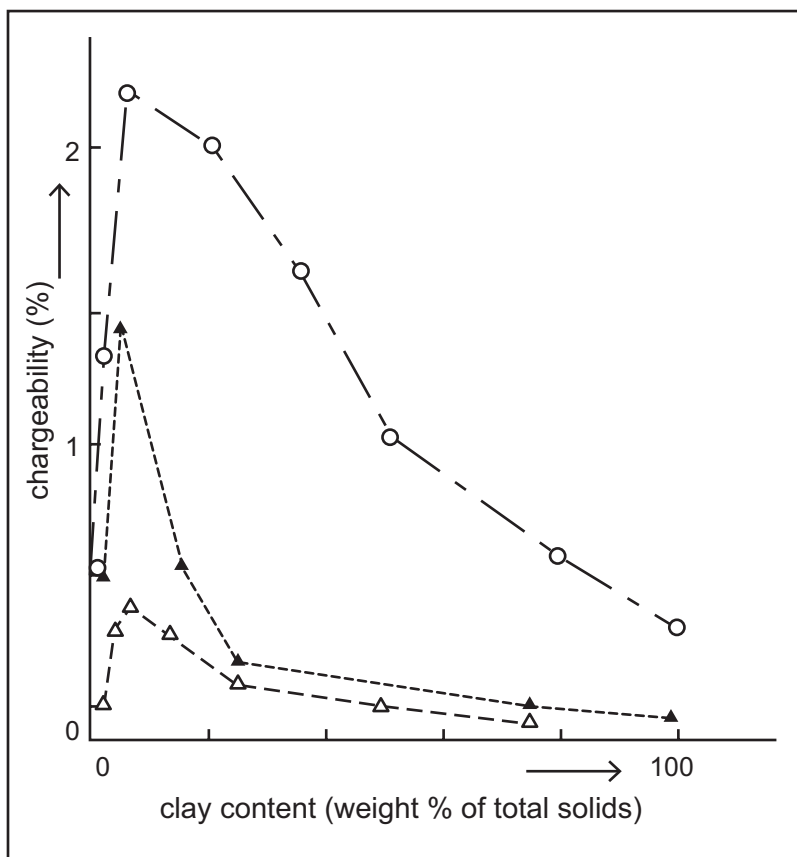


FIGURE 7.11A Variation of chargeability with clay content for three different clay minerals: C_w , unspecified; o, montmorillonite; ▲, bentonite; △, kaolin.

Source: Ogilvy and Kuzmina (1972).

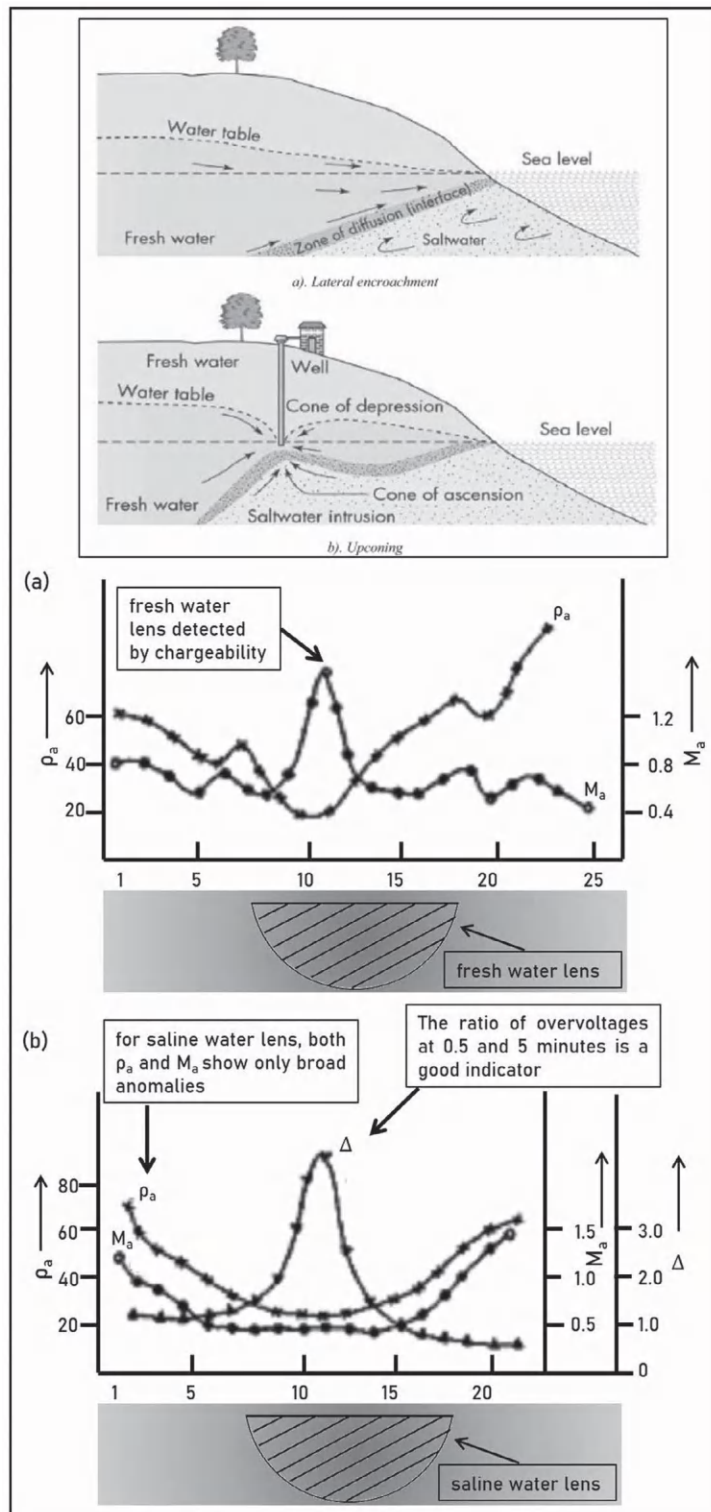


FIGURE 7.11B IP effect over freshwater and saltwater pockets.

Source: Ogilvy and Kuzmina (1972).

7.11.6 MINERAL EXPLORATION HIGH-PRODUCTIVITY SURVEYS OF MASSIVE AND DISSEMINATED SULPHIDE ORE BODY DETECTION

Unlike the electrical resistivity signature, very few materials produce a strong IP response. For natural soils, clay (membrane potential) and mineralized rocks (electrode potential) were characterized by a high chargeability. However, limestones and sandstones were characterized by a low chargeability. Figure 3, Page 2 in Reynolds (2011) shows low resistivity and high chargeability over a massive and disseminated sulphide ore body. The apparent resistivity and chargeability (or frequency effect or metal factor) data are carefully processed (inverted) using a specialist software. The figure in Reynolds (2011) shows resistivity and chargeability models of the ground resulting from the inversion of time domain IP data that were collected over a zone of suspected porphyry copper deposits, which typically have low resistivity and high chargeability (Reynolds, 2011).

The survey area is characterized by clay deposits accompanied by the presence of red mudstones which serve as the country rocks for Mn mineralization beneath the limestone (Wahyu Srigutomo et al., 2016). Both apparent resistivity and chargeability data for all lines were tabulated in pseudo-sections, which were then inverted numerically using the RES2DINV program (Loke and Barker, 1996), which transformed the pseudosections into models of true resistivity and chargeability as a function of lateral and vertical (2-D) distances (Wahyu Srigutomo et al., 2016). Resistivity values range between 1 and 100 Ω -m, which is considerably low from overall perspective. This feature may be attributed to the presence of manganese ore and/or clay matrix or rocks with low bulk resistivity. The rocks with manganese ore content are interpreted to have resistivity <5 Ω -m. However, to strengthen the reliability of the results, the resistivity sections are compared with the chargeability sections; those who have an intersection area between low resistivity and high chargeability (>10 msec) are considered to be the prospective zones of the manganese ore.

7.11.7 MIP SURVEY OVER TECK'S HIGHLAND VALLEY IN BRITISH COLUMBIA, CANADA

MIP survey was conducted over Teck's Highland Valley JA porphyry copper deposit in British Columbia, Canada. The area is known for having a conductive overburden that has made EIP surveys difficult (Chubak and Woods, 2013; Graden, 2013). Supracon's Jessy Deep High Temperature SQUIDS were used for both the survey measurements and the reference station. A base frequency of 0.25 Hz was used, with a 50% duty cycle. After referencing and stacking, the amplitudes were averaged over the time period from 442 ms to 992 ms. The figure in Chubak and Woods (2013; Graden, 2013) shows the MIP results plotted with the approximate outline of $>0.2\%$ Cu.

7.11.8 GEOPHYSICAL INVESTIGATION FOR BASE METAL MINERALIZATION IN KAROI-RAJPURA AREA, BHILWARA DISTRICT, RAJASTHAN, BY THE GEOLOGICAL SURVEY OF INDIA (GSI)

Strong geophysical anomalies were observed in the northern, southern, central, and eastern parts of the study area (Naskar et al., 2018). The low-resistivity zones (200 Ω -m) and high-chargeability zones (>24 mV/V) are situated in the southern and eastern parts of the investigated area (Naskar et al., 2018). However, to strengthen the reliability of the results, the resistivity map is compared with the chargeability map; those who have an intersection area between low resistivity and high chargeability (>24 mV/V) are considered to be the prospective zones of the base metal mineralization. However, these low-resistivity zones may have been influenced by the presence of clay or weathered soil. In this case, the high-chargeability zones will help in confirming the prospective zones caused by base metal mineralization. Recent deep drilling by Hindustan Zinc Limited (HZL) in this area has intersected rich ore lodes, thereby confirming the depth-wise continuity of ore zones at some places.

7.11.9 GEOPHYSICAL INVESTIGATION FOR COPPER AND GOLD MINERALIZATION AROUND THE DHANI–BASRI AND TODI KA BAS AREAS, DAUSA AND ALWAR DISTRICTS, RAJASTHAN, BY THE GSI

An IP survey has delineated several anomaly zones, where strike length varies from 400 m to 1,500 m. Chargeability values of zone I, zone II, and zone III range from 16 to 22 mV/V (Naskar and Kulkarni, 2019). Zone I, supported with low SP value of -100 to -200 mV, corroborates well with the Dhani–Basri known deposit. The chargeability values of zone IV, zone V, zone VI, and zone VII range from 15 to 24 mV/V, while the peak chargeability value of zone V is observed as 30 mV/V, over a background value of 10 mV/V. Chargeability-cum-resistivity pseudosections have shown depth persistence of the inferred conductors based on the contour pattern. These chargeability zones are associated with low SP, low resistivity, and moderate magnetic responses. Further, Bouguer anomaly profiles reflect a region where borehole intersected a mineralized zone (Figure 7.12).

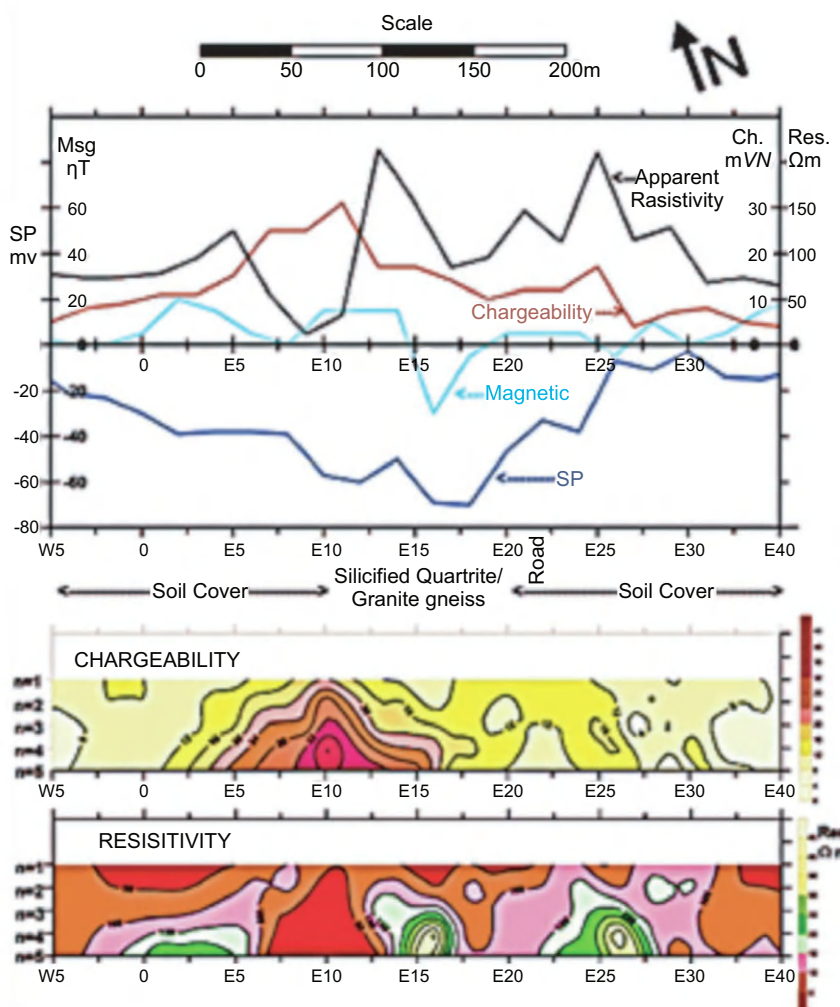


FIGURE 7.12 SP, magnetic (VF), IP (TD) chargeability, apparent resistivity profiles, along with chargeability and resistivity pseudosections along traverse 0, north of the Dhani–Basri area, Dausa and Alwar Districts, Rajasthan.

Source: Naskar et al. (2019).

7.11.10 GEOPHYSICAL INVESTIGATION FOR BASE METAL MINERALIZATION IN THE EXTENSION AREA OF THE MAMANDUR POLYMETALS PROSPECT IN VILLUPURAM AND THIRUVANNAMALAI DISTRICTS OF TAMIL NADU, INDIA

A combination of magnetic, IP, and resistivity methods was attempted in Mamandur polymetals prospect area of the Villupuram and Thiruvannamalai Districts of Tamil Nadu for locating base metal occurrence (Naskar et al., 2021). An IP survey was carried out in the area using a gradient array with current electrode separation of 1,000 m and potential electrode separation of 10 m and $n = 8$, where n is represented by a separation parameter between the current and potential electrode pair. The chargeability image map (Naskar et al., 2021) of the Mettukuppam mineralized zone has the chargeability values ranging up to 25 mV/V over a background value of 10 mV/V. The high chargeability is bounded between traverse S500 in the south to N100 in the north in between stations E230 in the east and W250 in the west. There are moderate to low chargeability values on both sides from E230 to E650 in the east and W250 to W350 in the west, as well as south of traverse S500. The anomaly zone is mainly confined in the shear zone and trending almost N–S direction due to the base metal sulphide mineralization. An IP-cum-resistivity pseudodepth section has been prepared for the estimation of depth and depth persistence of the causative body. A dipole–dipole array with $a = 10$ m and $n = 1$ –8 has been used for preparing the pseudodepth section. Both apparent resistivity and chargeability data were tabulated in pseudosections, which were then inverted numerically using the RES2DINV program (Loke and Barker, 1996), which transformed the pseudosections into models of true resistivity and chargeability as a function of lateral and vertical (2-D) distances. The figure in Naskar et al. (2021) shows, respectively, the inverse models of chargeability and resistivity over traverses S100, S200, S300, and S400, each with a 450 m strike length from E230 to W220 in the Mettukuppam mineral zone of the Vadamamandur–Porasapattu mineral block.

The rocks with mineralization content are interpreted to have resistivity of $<68 \Omega\text{-m}$ associated with the extreme west high-chargeability zone in between station W135 and W180 in S100. The rocks with mineralization content are $<74 \Omega\text{-m}$ associated with extreme west high chargeability zone in between stations W115 and W250 in S200 (Naskar et al., 2021). The rocks with mineralization content are interpreted to have resistivity $<57 \Omega\text{-m}$ associated with the extreme west part of the high-chargeability zone in between stations W75 and W250 in S300 (Naskar et al., 2021). In S400, one high-chargeability zone is observed in between stations E135 and E70, which is corroborated with moderate resistivity. Chargeability values range between 11 and 16 mV/V. This feature may be attributed to the presence of sulphide mineralization (Naskar et al., 2021). However, to strengthen the reliability of the results, the resistivity sections are compared with the chargeability sections; those who have an intersection area between low resistivity and high chargeability are considered to be the prospective zones of polymetals mineralization. Depth persistence of causative sources continues in each case beyond 20 m. A misfit between interpreted chargeability and resistivity is acceptably minimum to give the best inverted model.

7.11.11 GEOPHYSICAL INVESTIGATION OF THE Pb-Zn DEPOSIT OF LONTZEN–POPPELSBERG, BELGIUM

The drillhole information from the Lontzen–Poppelsberg site has demonstrated three ore bodies and has allowed the estimation of the extension of the lodes, their dip, and the location at the ground surface (Evrard et al., 2018). The localization of the lodes makes them excellent targets for further exploration with geophysics. This deposit is classified as a Mississippi Valley–type (MVT) deposit. It consists mainly of Pb-Zn-Fe sulphides that display contrasting values in resistivity and chargeability with regard to the sedimentary host rocks. The dipole–dipole direct current (DC) resistivity and induced polarization (IP) profiles have been collected and inverted to successfully delineate the Pb-Zn mineralization and the geological structures. The IP models were converted into normalized

chargeability (chargeability/resistivity). It allowed for better targeting of the sulphides. Based on a depth-of-investigation (DOI) threshold equal to 0.1, the inverted resistivity models were constrained by the datasets, up to a depth limit of 50 to 100 m (depending on the profiles) (Evrard et al., 2018; geological map modified from Laloux et al., 2010).

7.11.12 CASE STUDY: AN EXAMPLE OF AN IP ANOMALY OVER THE DURKIN SHOOT NICKEL PROSPECT, KAMBALDA DOME, WESTERN AUSTRALIA

An example of an IP anomaly over the Durkin Shoot, Kambalda Dome, is shown in Williams et al. (2005) (200 ft dipole, 0.3 and 3 Hz). Metal factor and frequency effect anomalies were common to all lines over the mineralization. Little indication of the dip of the mineralization is evident, however. Present-day IP surveys use lower frequency than that used in the initial Kambalda surveys in an effort to obtain better ground penetration. A frequency of 0.125 Hz and a dipole length of 50 m or greater are typical.

7.11.13 GOLD EXPLORATION USING INDUCED POLARIZATION (IP) AND RESISTIVITY METHODS IN THE MEYDUK–LATALA AREA (KERMEN PROVINCE): CASE STUDY

Due to the rock outcrops in the north of Meyduk (Latala) and preliminary geological surveys, 11 rectangle arrays of chargeability and resistivity with current electrode spacing of 600 m, station interval of 20 m, and profile interval of 50 m were carried out. The relative position of the study area is shown in Figure 1a, page 2 in Yousefi and Hafizi (2017). The survey data were contoured by Geosoft software in order to determine the prone area (Yousefi and Hafizi, 2017). It is expected that ore areas contain high chargeability and coloured spectral from orange to purple in the contoured IP map and have low resistivity in the resistivity map according to copper mineralization (as malachite and azurite in the fractures) and, in some cases, propylitic, sericitic, and argillic alterations. However, this case is violated in some sections due to low mineralization and the presence of very thin veinlet into high-electrical-resistance rocks. Two main veins are obvious in these contoured maps but only focus on the vein that is located in the right of the maps. Dipole–dipole array for more accuracy is used in the next step. The profiles are almost perpendicular to the main vein. In order to determine this vein, nine parallel profiles with electrode spacing of 20 m and 40 m were carried out. The data-derived profiles were plotted using Res2dinv and Surfer software programs (Yousefi and Hafizi, 2017). The direction of the expected anomaly is simply visible in the 2-D chargeability and resistivity sections. The results show that the depth of anomalies begins from a few meters and continues to more than 100 m, which indicate a good choice of the electrode spacing and the length of expanding.

7.11.14 APPLICATION OF GEOPHYSICS FOR MINERAL EXPLORATION IN SMALL-SCALE MINING IN FAZENDA SANTANA, BRAZIL: CASE STUDY

Many of the world-class deposits in operation are close to exhaustion, and new discoveries of this kind have been increasingly rare (Laznicka, 2010; Seccatore et al., 2013). Therefore, small-scale mining is likely to play a bigger role in mineral production in the future. For this reason, it has become increasingly important to study methods and technologies for small-scale mining. Tichauer et al. (2014) developed the methodology for relying on geophysics for mineral exploration in small-scale mining. This methodology has been applied to a small manganese deposit at Fazenda Santana, in the town of Itapira, Sao Paulo State, Brazil.

The methods used in the geophysical surveying were those of resistivity and induced polarization, and data were obtained through two complementary techniques, vertical electrical sounding (VES), with Schlumberger array, and electrical profiling, with dipole–dipole array. While VES surveys produce vertical and relatively deeper data, profiling generates information of layers in

sections and relatively lower depths (Braga, 2001; US EPA, 2014). These methods and techniques were chosen based on relatively low cost, ease and speed of mobilization and application, good response to low depths, and efficiency to individualize manganese layers. The Schlumberger array is the most widely used in VES because of the quality of field data, ease and speed of execution, and low sensitivity to lateral variations of resistivity and natural or artificial noises that exist underground, such as natural soil currents, interference from power lines, etc. Investigations by electrical profiling are conducted along lines, and the results are related to each other to generate sections showing the depths of subsurface layers. The dipole–dipole array was used due to the detail and precision of the results obtained. VES surveys were carried out on four points and horizontal profiling along five lines, as shown in Tichauer et al. (2014).

These surveys showed a layer with low resistivity ($<200 \Omega\text{-m}$) and high chargeability ($>21 \text{ mV/V}$) that differ from the others and interpreted as an occurrence of manganese. The results of VES indicate, as in Table 7.4, that the mineralized layer can exceed 10 m thick, and parts of the deposit can be found over 40 m deep. The five electrical profiling surveys were carried out in the NW–SE direction, transverse to the deposit axis. The resistivity sections generated are shown in Tichauer et al. (2014). The mineralized layer, represented by resistivity of $<200 \Omega\text{-m}$, is illustrated with shades of green and blue. Chargeability data confirm the resistivity information. Resistivity sections indicate that a significant part of the deposit should be found in depths not exceeding 15 m, and that the layer can exceed 10 m thick. Section 106 shows an anomaly, which electrical characteristics suggest an occurrence of a shallow mineralized zone. The data obtained through VES and profiling strengthened the evidence of occurrence of an economically viable manganese deposit, enabled preliminary modelling of the deposit, and resulted in a lean, cost-effective drilling plan for Fazenda Santana.

TABLE 7.4**Depth and Thickness of Surface Layers Obtained through VES Surveys**

Survey		From (m)	To (m)	Rock	Inferences
VES 105	S-01	0	0.6	Soil	
	S-01	0.6	1.1	Soil	
	S-01	1.1	10.0	Unconsolidated rock	
	S-01	10.0	35.0	Unconsolidated rock	
	S-01	35.0		Manganese	End of mineralized
VES 107	S-02	0	0.7	Soil	
	S-02	0.7	9.8	Soil	
	S-02	9.8	28.0	Unconsolidated rock	
	S-02	28.0		Manganese	End of mineralized
VES 108	S-03	0	0.6	Soil	
	S-03	0.6	13.5	Soil	
	S-03	13.5	18.6	Unconsolidated rock	
	S-03	18.6	29.0	Manganese	29–18.6 m or 10.4 m thick mineralized layer
VES 109	S-04	0	0.5	Soil	
	S-04	0.5	6.6	Soil	
	S-04	6.6	42.9	Unconsolidated rock	
	S-04	42.9	55.0	Manganese	Part of the body can be 42.9 m deep, 55–42.9 m or 12.1 m thick mineralized layer
	S-04	55.0		Bedrock	

Source: Tichauer et al. (2014).



FIGURE 7.13 View of the geological model exposed on the surface with conceptual benches.

Source: Tichauer et al. (2014).

The combination of geophysics data with the conceptual deposit model resulted in the preliminary model of the manganese deposit at Fazenda Santana seen in Tichauer et al. (2014) and Figure 7.13.

With the data obtained, a plan of exploratory drilling was represented in Tichauer et al. (2014). The 1,660 m of drilling should be according to the following parameters. Phase 1: net of approximately $150\text{ m} \times 50\text{ m}$, on parallel lines coincident with the lines of the horizontal profiling, with 10 holes and estimated total length of 780 m of drilling. Phase 2: ‘infill’ net of approximately $75\text{ m} \times 50\text{ m}$, with parallel lines and interspersed lines of phase 1, with 12 holes and estimated total length of 880 m of drilling. All drilling must be executed at vertical dip between 60 and 100 m, according to the conceptual morphology of the geological model.

REFERENCES

- Aiken, C.L.V. and Hastings, D., 1973. Numerical modelling of induced polarisation. 11th International Symposium of Computer Applications in the Mining Industry, Amsterdam, Netherlands, College of mines, University of Arizona.
- Binley, A., 2015. Tools and techniques: Electrical methods. In: *Treatise on Geophysics*. Vol. 11. Amsterdam, Netherlands: Elsevier B.V., pp. 233–259.
- Bodmer, R., Ward, S.H. and Morrison, H.F., 1968. On induced electrical polarization and groundwater. *Geophysics*, vol. 33, no. 5, pp. 805–821.
- Borner, F.D., Schopper, J.R. and Weller, A., 1996. Evaluation of transport and storage properties in the soil and groundwater zone from induced polarization measurements. *Geophysical Prospecting, European Association of Geoscientists & Engineers (EAGE)*, vol. 44, pp. 583–602.
- Braga, A.C.O., 2001. Metodos Geoeletricos Aplicados: Modulo: Hidrogeologia, IGCE—unesp, Rio Claro, p. 91.
- Cahyna, F.O., Mazak and Venhodova, D., 1990. Determination of the extent of cyanide contamination by surface geoelectrical methods. In: Ward, S.H. (ed), *Geotechnical and Environmental Geophysics, vol. II, Environmental and Groundwater, Invest. Geophys.* Vol. 5. Tulsa, OK: Society of Exploration Geophysicists. pp. 97–99.
- Chen, J., Haber, E. and Oldenburg, D.W., 2002. Three dimensional numerical modelling and inversion of magnetometric resistivity data. *Geophysical Journal International (GJI)*, Oxford Academic, Publisher Oxford University Press, vol. 149, pp. 679–697.
- Chubak, G. and Woods, D., 2013. Magnetic induced polarization—using new technology for greater detection

- capability of deep and elusive mineralization. 23rd International Geophysical Conference and Exhibition, 11–14 August 2013. Melbourne, Australia.
- Coggon, J.A., 1971. Electromagnetic and electrical modelling by the finite element method. *Geophysics*, vol. 36, pp. 137–155.
- Coggon, J.A., 1973. A comparison of IP electrode arrays. *Geophysics*, vol. 38, pp. 737–761.
- Cole, K.S. and Cole, R.H., 1941. Dispersion and adsorption in dielectrics. Alternating current field. *Journal of Chemical Physics*, Published by American Institute of Physics (AIP), vol. 9, p. 341.
- Dahlin, T. and Zhou B., 2006. Multiple-gradient array measurements for multichannel 2D resistivity imaging. *Near Surface Geophysics*, vol. 4, pp. 113–123.
- Evrard, M., Gael, D., Thomas, H., Michel, C., Olivier, F., Eric, P. and Frederic, N., 2018. Geophysical investigation of the Pb-Zn deposit of Lontzen-Poppelsberg, Belgium. *Minerals*, vol. 8, no. 6, p. 233. DOI: <https://doi.org/10.3390/min8060233>.
- Frederic, N. and Robin, T., (academic year) 2016–2017. Application of the time-domain induced polarization technique to characterize the Orman gold prospect in Mondulkiri Province, Cambodia. University of Liege, Graduation Studies conducted for obtaining the Master's degree in Mining and Geological Engineering by Thibaut Robin, p. 134.
- Frische, R.H. and Von Buttlar, H., 1957. A theoretical study of induced electrical polarization. *Geophysics*, vol. 22, no. 3, pp. 688–706.
- Graden, R.W., 2013. Technical report teck Highland Valley Copper Highland Valley, British Columbia, Canada to: Teck resources limited, NI 43–101. Presented in 23rd International Geophysical Conference and Exhibition, 11–14 August 2013. Melbourne, Australia.
- Iliceto, V., Santarato, G. and Veronese, S., 1982. An approach to the identification of fine sediments by induced polarization laboratory measurements. *Geophysical Prospecting, European Association of Geoscientists & Engineers (EAGE)*, vol. 30, pp. 331–347.
- Kearey, P., Brooks, M. and Hill, I., 2002. *An Introduction to Geophysical Exploration*. 3rd edn. Hoboken, NJ: Blackwell Science Ltd.
- Kemna, A., 2000. Tomographic inversion of complex resistivity. Ruhr-University at Bochum. Vol. 169. Available at www2.geo.uni-bonn.de/members/kemna/Kemna{}_2000{}_PhD{}_thesis.pdf.
- Kemna, A., Binley, A., Cassiani, G., Niederleithinger, E., Revil, A., Slater, L., Kenneth, H.W., Orozco, A.F., Haegel, F.H., Hordt, A., Kruschwitz, S., Leroux, V., Titov, K. and Zimmermann, E., 2012. An overview of the spectral induced polarization method for near-surface applications. *Near Surface Geophysics*, vol. 10, pp. 453–468.
- Klein, J.D. and Sill, W.R., 1982. Electrical properties of artificial clay-bearing sandstone. *Geophysics*, vol. 47, no. 11, pp. 1593–1605.
- Knoll, M.D., Knight, R.J. and Brown, E., 1994. Can accurate estimates of permeability be obtained from measurements of dielectric properties? Symposium on the Application of Geophysics to Engineering and Environmental Problems (SAGEEP), Boston, MA, pp. 25–35.
- Kuzmina, E.N. and Ogilvy, A.A., 1965. On the possibility of using the induced polarization method to study groundwater. *Razvedochnaya i Promyslovaya Geofizika*, Published by Gosgortekhzdat in Moscow, no. 9, pp. 47–69.
- Laloux, M., Geukens, F., Ghysel, P. and Hance, L.C., 2010. *Géologique de Wallonie Henri-Chapelle-Raeren 43/1–2*. Bruxelles, Belgium: Service géologique de Belgique, p. 201.
- Laznicka, P., 2010. *Giant Metallic Deposits*. 2nd edn. Berlin: Springer-Verlag, p. 949.
- Loke, D.M., 1999. *Electrical Imaging Surveys for Environmental and Engineering Studies*. Cangkat Minden Lorong, vol. 63, p. 6574525.
- Loke, M.H., 2015. Electrical imaging survey for environmental and engineering studies. A Practical Guide to 2D and 3D Surveys. vol. 187. Available at: [http://seisweb.oma.be/bibadmin/uploads/pdf/COURSE-NOTES.pdf\[%\]5Cnwww.geotomosoft.com](http://seisweb.oma.be/bibadmin/uploads/pdf/COURSE-NOTES.pdf[%]5Cnwww.geotomosoft.com).
- Loke, M.H. and Barker, R.D., 1996. Rapid least squares inversion of apparent resistivity pseudosections by a quasi-Newton method. *Geophysical Prospecting*, vol. 44, pp. 131–152.
- Luo, Y. and Zhang, G., 1998. *Theory and Application of Spectral Induced Polarization*. Tulsa, OK: Society of Exploration Geophysicists.
- Marshall, D.J. and Madden, T.R., 1959. Induced polarisation, a study of its causes. *Geophysics*, vol. 24, pp. 790–816.
- Martinho, E., Abreu, M.M., Oliveira, A., Alegria, F. and Pam Pulha, M.E., 2013. Monitoring of bio-physicochemical changes in a silty clay soil contaminated with LNAPLs. *Geoderma*, vol. 197–198, pp. 108–116.

- Naskar, D.C. and Kulkarni, A.V., 2019. Geophysical investigation for copper and gold mineralization around Dhani-Basri and Todi ka Bas area, Dausa and Alwar districts, Rajasthan. *Journal of Indian Geophysical Union*, vol. 23, no. 2, pp. 120–133.
- Naskar, D.C., Kulkarni, A.V. and Lakshmana, M., 2018. Geophysical investigation for base metal mineralization in Karoi-Rajpura area, Bhilwara district, Rajasthan. *Journal of Indian Geophysical Union (JIGU)*, Hyderabad, vol. 22, no. 4, pp. 379–388.
- Naskar, D.C., Virendra, K. and Muktibodh, P., 2021. Geophysical investigation for base metal mineralization in extension area of Mamandur polymetals prospect in Villupuram and Thiruvannamalai districts of Tamil Nadu. *Journal of Indian Geophysical Union*, vol. 25, no. 2, pp. 30–44.
- Ogilvy, A.A. and Kuzmina, E., 1972. Hydrogeologic and engineering-geologic possibilities for employing the method of induced potentials. *Geophysics*, vol. 37, no. 5, pp. 839–861.
- Olhoeft, G.R., 1984. Clay-organic interactions measured with complex resistivity. The 54th Annual International Meeting of the Society of Exploration Geophysicists (SEG), Atlanta, GA, Expanded Abstracts, pp. 356–358.
- Olhoeft, G.R., 1985. Low-frequency electrical properties. *Geophysics*, vol. 50, pp. 2492–2503.
- Olhoeft, G.R., 1992. Geophysical detection of hydrocarbon and organic chemical contamination: Proc. Symp. On the Application of Geophysics to Engineering and Environmental Problems, pp. 587–594.
- Parasnis, D.S., 1966. *Principles of Applied Geophysics*. 5th edn. London: Chapman & Hall.
- Parasnis, D.S., 1973. *Mining Geophysics*. Amsterdam, London and New York: Elsevier Publishing Company.
- Pelton, W.H., Ward, S.H., Hallof, P.G. and Nelson, P.H., 1978. Mineral discrimination and removal of inductive coupling with multifrequency IP. *Geophysics*, vol. 43, pp. 588–609.
- Reynolds, J.M., 2011. *An Introduction to Applied and Environmental Geophysics*. 2nd edn. Chichester: John Wiley & Sons Ltd. p. 712.
- Seara, J.L. and Granda, A., 1987. Interpretation of IP time-domain/resistivity sounding for delineating sea-water intrusions in some coastal areas of the northeast of Spain. *Geoexploration*, vol. 24, pp. 153–161.
- Seccatore, J., Marin, T., De Tomi, G. and Veiga, M., 2013. A practical approach for the management of resources and reserves in small-scale mining. *Journal of Cleaner Production*, p. 6, ISSN 0959-6526, Elsevier.
- Seigel, H.O., 1959. Mathematical formulation and type curves for induced polarization. *Geophysics*, vol. 24, pp. 547–565, DOI: 10.1190/1.1438625.
- Seigel, H.O., 1974. The Magnetic Induced Polarization (MIP) method. *Geophysics*, Dordrecht, Netherlands, vol. 30, pp. 332–339.
- Seigel, H.O. and Howland-Rose, A.W., 1983. *The Magnetic Induced Polarization Method: Developments in Geophysical Exploration Methods*. 4th edn. Dordrecht, Netherlands: Springer, pp. 65–69.
- Seigel, H.O. and Howland-Rose, A.W., 1990. Magnetic induced-polarization method. In: Fink, J.B., Sternberg, B.K., McAlister, E.O., Wieduwilt, W.G. (eds), *Induced Polarization: Applications and Case Histories*. Tulsa, OK: Society of Exploration Geophysicists. pp. 23–56.
- Slater, L.D. and Lesmes, D., 2002. IP interpretation in environmental investigations. *Geophysics*, vol. 67, no. 1, pp. 77–88, DOI: 10.1190/1.1451353.
- Slater, L.D. and Sandberg, S.K., 2000. Resistivity and IP monitoring of salt transport under natural hydraulic gradients. *Geophysics*, vol. 65, pp. 408–420.
- Slater, L., Dimitrios, N., Yves, R.P. and Susan, H., 2007. Pore-scale spectral induced polarization signatures associated with FeS bio-mineral transformations. *Geophysical Research Letters*, vol. 34, L21404, DOI: 10.1029/2007GL031840.
- Srigutomo, W., Trimadana and Prihandhanu, M.P., 2016. 2D resistivity and induced polarization measurement for manganese ore exploration. 6th Asian Physics Symposium, Journal of Physics, Conference Series 739, IOP Publishing (2016) 012138, DOI: 10.1088/1742-6596/739/1/012138.
- Sternberg, B.K. and Oehler, D.Z., 1990. Induced polarisation in hydrocarbon surveys: Arkoma basin case histories. In: Ward, S.H. (ed), *Induced Polarisation: Applications and Case Histories*. Vol. 4. Tulsa, OK: Society of Exploration Geophysicists.
- Sturrock, J.T., Lesmes, D. and Morgan, F.D., 1999. Permeability estimation using spectral induced polarization measurements. Proceedings of the Symposium on the Application of Geophysics to Engineering and Environmental Problems, Portland, Oregon, pp. 409–415.
- Sumner, J.S., 1976. *Principles of Induced Polarisation for Geophysical Exploration*. Amsterdam: Elsevier Scientific Publishing Company.
- Telford, W.M., Geldart, L.P. and Sheriff, R.E., 1990. *Applied Geophysics*. Cambridge: Cambridge University Press.

- Tichauer, R.M., Marin, T., Martins, A.C., de Tomi, G. and Elis, V.R., 2014. Application of geophysics for mineral exploration in small scale mining. Conference: VIII CBMINA, at: Belo Horizonte MG. Vol. 1.
- Towel, J.N., Anderson, R.G., Pelton, W.H., Olhoeft, G.R. and LaBrecque, D., 1985. Direct detection of hydrocarbon contaminants using induced polarisation method. SEG Meeting, SEG International Exposition and Annual Meeting, Washington, DC, pp. 145–147.
- US EPA, 2014. Environmental Geophysics. U.S. Environmental Protection Agency, U.S. Environmental Protection Agency's (EPA) Work on Environmental Geophysics, United States. Consultado em 25/01/2014.
- Vacquier, V., Holmes, C.R., Kintzinger, P.R. and Lavergne, M., 1957. Prospecting for groundwater by induced electrical polarisation. *Geophysics*, vol. 23, pp. 660–687.
- Vanhala, H., Soininen, H. and Kukkonen, I., 1992. Detecting organic chemical contaminants by spectral-induced polarization method in glacial till environment. *Geophysics*, vol. 57, pp. 1014–1017.
- Vinegar, H.J. and Waxman, M.H., 1984. Induced polarization of shaly sands. *Geophysics*, vol. 49, pp. 1267–1287.
- Wait, J.R., 1959a. Overvoltage research and geophysical application. In: *International Series of Monographs on Earth Sciences*. Vol. 4. Oxford, England: Pergomon Press, pp. 22–83.
- Wait, J.R., 1959b. A phenomenological theory of overvoltage for metallic particles. In: *Overvoltage Research and Geophysical Application*. Vol. 4. Oxford, England: Pergomon Press, pp. 22–28.
- Wait, J.R., 1959c. The variable—frequency method. In: *Overvoltage Research and Geophysical Application*. Vol. 4. Oxford, England: Pergomon Press, pp. 29–49.
- Wightman, W.E., Jalinoos, F., Sirles, P. and Hanna, K., 2003. *Application of Geophysical Methods to Highway Related Problems*. Lakewood, CO: Federal Highway Administration, Central Federal Lands Highway Division, Publication No. FHWA-IF-04–021, September 2003.
- Yousefi, M.R. and Hafizi, M.K., 2017. Gold exploration using induced polarization and resistivity methods in Meyduk-Latala area. 79th EAGE Conference & Exhibition 2017—Student Programme Paris, France.
- Zohdy, A.A.R., 1974a. Use of Dar Zarrouk curves in the interpretation of vertical electrical sounding data: U.S. Geol. Survey Bull. 1313-D, p. 41.
- Zohdy, A.A.R., 1974b. A computer program for the calculation of Schlumberger sounding curves by convolution: Available only from U.S. Dept. Commerce Natl. Tech. Inf. Service, Springfield, Va. 22161, as U.S. Geol. Survey Rept. USGS-GD-74–010, PB-232 056.

8 Electromagnetic (EM) Methods

8.1 INTRODUCTION

The pioneering work of the Schlumberger brothers stimulated the creation and development of the Russian school of EM geophysicists, which produced a significant number of distinguished geophysicists and led to major geologic discoveries in the Ural Mountains and eastern Siberia in the 1940s and 1950s. It is a documented fact that electromagnetic geophysical methods are based on the study of the propagation of electric currents and subsequent generated electromagnetic fields inside the subsurface of the Earth. There are two principal mechanisms of electromagnetic methods: one is based on direct current (DC), and another one works on a transient electromagnetic field. The electromagnetic techniques have the broadest range of different instrumental systems. They can be classified into either time domain (TEM) or frequency domain (FEM) systems. The FEM uses one or more frequencies, while TEM measurements are based on a function of time. Depending upon the use of the sources, EM methods can be either passive, utilizing natural ground signals (e.g. magnetotellurics) as a very proactive trait of EM methodology, where an artificial transmitter is used either in the near field (as in ground conductivity meters) or in the far field (using remote high-powered military and civil radio transmitters, as in the case of very-low-frequency [VLF] and RMT [radio magnetotelluric] methods). The aim of the measurements is the same as we do for geoelectric geophysical surveys using direct current (DC) as the source for conducting the survey. It is evident that the determination of resistivity is a function of depth or a function of profile distance. The main advantage of the EM methods is that these techniques do not require direct contact with the ground, as in the case of DC electrical methods. Therefore, the EM measurements can be carried out in a faster way than the DC measurements.

The number of different modifications of EM methods is extremely large because we have the possibility of using different types of transmitters and measuring the different components of EM fields in a variety of receivers. The behaviour of an EM field is controlled by three primary properties of the medium—the electrical conductivity, the dielectric permittivity, and the magnetic susceptibility—as well as a number of other physical phenomena. Nonetheless, the electrical conductivity of the material or media is the most useful of all physical properties that characterize rocks and minerals in geophysical exploration with relatively low frequency. For high-frequency methods such as ground-penetrating radar (GPR), dielectric permittivity is the most important property. It is known that the electrical conductivity in the Earth varies by many orders of magnitude; in contrast, the other physical properties used in geophysical exploration (acoustic wave speed, density, magnetic susceptibility) vary over quite limited ranges. Many different parameters affect electrical conductivity, the most important being the presence of water or hydrocarbons in a rock, ground-water salinity, amounts of electronically conducting minerals in a rock, temperature, and pressure. All these factors make electrical conductivity an obvious tracer for economic mineral deposits of various types.

Electromagnetic induction methods can be used for ground, airborne, sea-bottom, and borehole observations. Generally, the surface electromagnetic field uses surface loops and a receiver or deploys a downhole tool lowered into a borehole and acts on the body of mineralization, whilst the regional electromagnetic field uses airborne surveillance, either a fixed-wing aircraft or a helicopter equipped with electromagnetic tools. Interestingly, both methods can map 3-D perspectives of mineral bodies, deep in the Earth, and guide further exploratory drilling for verification. The range of EM applications is large. These methods are applied in the mining industry; in oil and gas exploration; in geotechnical, engineering, groundwater, and environmental geophysics; and in regional and crustal tectonic studies. The two main applications of electrical and EM methods are sounding, mostly for

TABLE 8.1
The Range of Applications for EM Surveying (Independent of Instrument Type)

Methodology	Applications
Electromagnetic (EM) method	Mineral exploration
	Mineral resource evaluation
	Groundwater surveys
	Mapping contaminant plumes
	Geothermal resource investigations
	Contaminated land mapping
	Landfill surveys
	Detection of geological and artificial cavities
	Location of geological faults, etc.
	Geological mapping
	Permafrost mapping, etc.

petroleum and other stratigraphic studies, and searching for discrete conductors in base metal exploration. It is dependent upon the type of equipment but can be categorized as listed in Table 8.1.

The first successful application of electrical and EM methods was in exploring for highly conductive metal ores. The methods were applied in the exploration for massive sulphide ore bodies and disseminated metal ores. Even with the development of other applications in more recent years, the use of EM methods in the search for metallic ores remains one of the most important commercial applications. The fact is that in the big picture of expenditures in exploration geophysics, EM methods are still small compared to seismic methods, and within EM methods, the search for minerals has held the dominant position.

8.2 PRINCIPLES OF EM SURVEYING

Electromagnetic (EM) methods make use of the response of the ground to the propagation of the electromagnetic fields, which are composed of alternating electric intensity and magnetic force. All entities (*E, D, H, B, J*) of the electrical and magnetic fields that define the electromagnetic wave are expressible in terms of vectors having different units:

- E* is the electrical field, in V/m.
- D* is the dielectric displacement, in coulomb/m².
- H* is the magnetic field intensity, in A/m.
- B* is the magnetic induction, in tesla.

It is observed that the entire electromagnetic phenomena obey the following four Maxwell equations.

8.2.1 FARADAY’S LAW

The Maxwell–Faraday equation states that a time-varying magnetic field always accompanies a spatially varying (also possibly time-varying), non-conservative electric field, and vice versa (Kearey and Brooks, 2002). The Maxwell–Faraday equation is

$$\nabla \times \mathbf{E} = - \frac{\partial \mathbf{B}}{\partial t}$$

8.2.2 AMPERE'S LAW

$$\nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t}$$

It shows us how an electric current and/or a time-varying electric field generates a magnetic field (Kearey and Brooks, 2002).

$$\text{div } \mathbf{B} = 0$$

It infers that lines of magnetic induction are continuous and there are no single magnetic poles.

$$\text{div } \mathbf{D} = q$$

It infers that electrical fields can begin and end on electrical charges.

By using the following subsidiary equations

$$\mathbf{D} = \epsilon \mathbf{E} \quad \mathbf{B} = \mu \mathbf{H} \quad \mathbf{J} = \sigma \mathbf{E}$$

where J is the electrical current density in A/m², q is the electric charge in coulomb/m³, ϵ is the electrical permittivity, μ is the magnetic permeability, σ is the electrical conductivity, and the four Maxwell equations, the electromagnetic wave equation can be derived. Such waves, with low attenuation, and their relationship, are shown in the figure in Kearey and Brooks (2002).

8.3 ELECTROMAGNETIC INDUCTION

Electromagnetic (EM) surveys carried out at frequencies below 50 kHz are based on the principle of electromagnetic induction. An alternating magnetic field in a coil or cable induces electric currents in a conductor. Since the conductivity of rocks and soils is too poor to permit significant induction currents when a good conductor is present, a system of *eddy currents* is set up. In turn, the eddy currents produce secondary magnetic fields that are superposed on the primary field and can be measured at the ground surface (Figure 8.1a).

Suppose that a low-frequency plane wave propagates along the vertical z -axis. The displacement current is now negligible compared to the conduction current, and the Maxwell equation for B field becomes (see derivation in Chapter 9)

$$\frac{\partial^2 \mathbf{B}}{\partial z^2} = \mu_0 \sigma \frac{\partial \mathbf{B}}{\partial t} \quad (8.1)$$

In the preceding equation, the magnetic field has components B_x and B_y . The form of this equation is reminiscent of the one-dimensional equation of *diffusion* or *heat conduction*, whose solution describes how the temperature changes with time and position when a fluctuating temperature acts on the surface. By analogy, the solution of eq. 8.1 for the components B_x or B_y of an alternating magnetic field with angular frequency ω ($= 2\pi f$) in a conductor with conductivity σ is

$$B_{x,y}(z, t) = B_0 e^{-\frac{z}{d}} \cos\left(\omega t - \frac{z}{d}\right) \quad (8.2)$$

$$\text{where } d \text{ is } \sqrt{\frac{2}{\mu_0 \sigma \omega}} \quad (8.3)$$

Here, d is called the *skin depth*. At this depth, the magnetic field is attenuated to e^{-1} ($\sim 37\%$) of its value outside the conductor. The skin depth is dependent on the conductivity of the body and

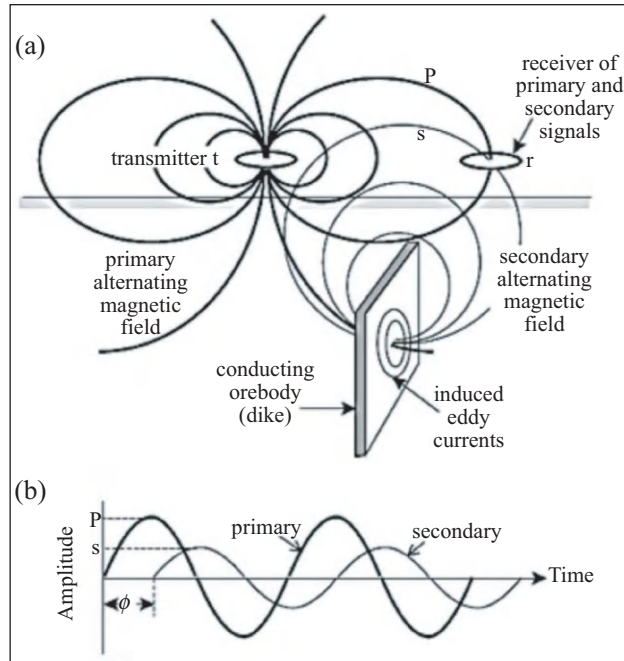


FIGURE 8.1 (a) Illustration of the primary and secondary fields in the horizontal loop induction method of electromagnetic exploration for shallow ore bodies. (b) Amplitudes and phases of the primary (p) and secondary (s) fields.

Source: (a) Burger et al., 2023; (b) [www.kau.edu.sa/Files/0003035/Subjects/EM\(1\).pdf](http://www.kau.edu.sa/Files/0003035/Subjects/EM(1).pdf).

the frequency of the field. The skin depth in normal ground ($\sigma \sim 10^{-3} \Omega^{-1} \text{m}^{-1}$) for a low-frequency alternating magnetic field ($f \sim 10^3$ Hz) is about 500 m, but in an ore body ($\sigma \sim 10^3 \Omega^{-1} \text{m}^{-1}$), it is only ~ 16 cm. The comparable figures for a high-frequency radar signal ($f \sim 10^9$ Hz) are 50 cm and 0.16 mm, respectively. It is important to note that the skin depth is not the maximum depth of penetration of the magnetic field. It helps to indicate how rapidly the field is attenuated, but the magnetic field is effective at depths that are many times the skin depth. However, it decays to $\sim 1\%$ at a depth $z = 5d$ and to $\sim 0.1\%$ at $z = 7d$, effectively limiting the practical depth of exploration with the induction method.

The many field methods of EM induction have a common principle. A coil or cable is used as a transmitter of the primary alternating magnetic field, while another coil serves as a receiver of both the primary signal and a secondary signal from the eddy currents induced in a conductor (Figure 8.1a). The magnetic field in the conductor experiences a phase shift equal to $(\frac{z}{d})$ (eq. 8.2) due to the conductivity. This results in a phase difference ϕ between the secondary and primary signals in the receiver (Figure 8.1b). It is observed that the exact theory of EM induction is complicated, even in a simple situation, but we can obtain a simple qualitative appreciation by applying some concepts from electrical circuit theory. We will use complex numbers, involving $i = \sqrt{-1}$.

Let the current systems in the transmitter, receiver, and conductor be represented by simple loops carrying currents I_t , I_r , and I_c , respectively. If the currents are sinusoidal, each has the form $I = I_0 e^{-i\omega t}$, so that $\frac{dI}{dt} = i\omega I$. Let the resistance of the conductor be R and its self-inductance be L . The voltage V_c in the conductor is composed of two parts: a resistive part due to the current I_c in the

resistance R being equal to $(I_c R)$, and an inductive part due to the change of current being equal to $L \frac{dI_c}{dt}$. Then complete voltage in the conductor can be expressed as:

$$V_c = I_c R + L \frac{dI_c}{dt} = I_c (R + i\omega L) \quad (8.4)$$

The voltage V_c is induced in the conductor by the changing current I_t in the transmitter circuit. Let the mutual inductance between transmitter and conductor be M_{tc} , then $V_c = -M_{tc} \frac{dI_t}{dt}$. Similarly, the transmitter current induces a primary voltage $V_p = -M_{tr} \frac{dI_t}{dt}$ in the receiver, in which the eddy currents in the conductor also induce a secondary voltage $V_s = -M_{cr} \frac{dI_c}{dt}$. Here, M_{tr} and M_{cr} are the mutual inductances between transmitter and receiver, and conductor and receiver, respectively.

We can express the following relationships that exist between the different voltages and currents:

$$V_p = -M_{tr} \frac{dI_t}{dt} = -i\omega M_{tr} I_t \quad (8.5)$$

$$V_s = -M_{cr} \frac{dI_c}{dt} = -i\omega M_{cr} I_c \quad (8.6)$$

$$V_c = -M_{tc} \frac{dI_t}{dt} = -i\omega M_{tc} I_t \quad (8.7)$$

Combining eq. 8.4 and eq. 8.7, we get

$$\frac{I_c}{I_t} = -\frac{i\omega M_{tc}}{R + i\omega L} = -\frac{i\omega M_{tc}}{R^2 + \omega^2 L^2} (R - i\omega L) \quad (8.8)$$

From eq. 8.5 and eq. 8.6, the ratio of V_s to V_p in the receiver is

$$\frac{V_s}{V_p} = \frac{M_{cr} I_c}{M_{tr} I_t} = -\frac{M_{tc} M_{cr}}{M_{tr}} \frac{(\omega^2 L + i\omega R)}{R^2 + \omega^2 L^2} \quad (8.9)$$

Which can be written in the form

$$\frac{V_s}{V_p} = \frac{M_{cr} I_c}{M_{tr} I_t} = -\frac{M_{tc} M_{cr}}{M_{tr} L} \left(\frac{\beta^2 + i\beta}{1 + \beta^2} \right) \quad (8.10)$$

where $\beta = \frac{\omega L}{R}$ is the response parameter of the conductor. The function in parentheses in eq. 8.10 is a complex number, so the voltage ratio (or *response* of the measuring system) can be written $P + iQ$. It should be noted that the real part P has the same phase as the primary signal and is called the *in-phase component* of the response. The imaginary part Q is 90° out of phase with the primary signal (i.e. if the primary signal is $\sim \cos \omega t$, the imaginary part is $\sim \sin \omega t = \cos(\omega t - \frac{\pi}{2})$; it is called the *quadrature component*).

8.4 EM INDUCTION SURVEYING

8.4.1 HORIZONTAL LOOP EM METHOD

The resemblance of the basic equations of EM induction to the diffusion equation $\frac{\partial T}{\partial t} = \kappa (\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2})$ (where T is the temperature and κ the thermal diffusivity) classifies the method as a *diffusive* one. Diffusive techniques—for example, the gravity, magnetic field, geothermal, and seismic surface wave methods—respond to a volumetric average of the specific physical parameter and do not show a fine detail of its distribution. It demonstrates that EM induction yields an average value of the electrical conductivity in a particular volume with the resolution better than that of potential field methods.

In the horizontal loop electromagnetic method (HLEM), commonly known also as *Slingram* or *Ronka*, the receiver and transmitter are coupled by a fixed cable about 30 to 100 m in length. During the survey, the receiver and transmitter should be kept at a constant separation while the pair is moved along a traverse of a suspected conductor (Figure 8.2a). In this field set-up, the cable supplies a direct signal that exactly cancels the primary signal at the receiver, leaving only the secondary field of the conductor. This is separated into the in-phase and quadrature components, which are expressed as percentages of the primary field and plotted against the position of the midpoint of the pair of coils (Figure 8.2b).

At the time of analyzing EM data, we ascertain if the in-phase and quadrature signals are zero far from the conductor and at the places where either the transmitter or the receiver passes over the conductor, then that enables us to trace the outline of a buried conductor to be charted. The signal

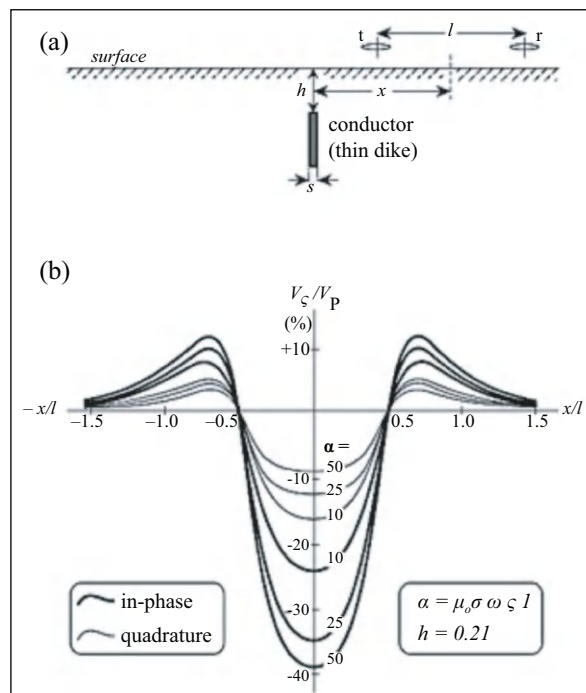


FIGURE 8.2 (a) Geometry of an HLEM profile across a thin vertical dyke. (b) In-phase and quadrature profiles over a dyke at depth $\frac{h}{l} = 0.2$ for some values of the dimensionless response parameter α .

Source: Burger et al. (2023).

risers to a positive peak on both sides and falls to a negative peak over the middle of the conductor. The peak-to-peak responses of the in-phase and quadrature components depend on the quality of the conductor, which is expressed by a response parameter such as β in eq. 8.10. A suitable function is the dimensionless parameter $\alpha = \mu_0 \sigma \omega s l$, which contains the conductivity σ and width s of the conductor, as well as the coil spacing l and frequency ω of the EM system. The systematic variation of the response curves with the value of α (Figure 8.2b) allows interpretation of the quality of the conductor. The variation of in-phase and quadrature signals over a conducting ore body can be modelled experimentally on a smaller scale in the laboratory. The smaller values of s and l are compensated by the larger values of σ and ω to give the same response parameter α . We generate model response curves for different α , which are then directly applied to the interpretation of real conductors measured in the field.

It is reported that integrated geophysical studies including EM methods were used in India by mining companies and national organizations, like the Geological Survey of India (GSI) and the National Geophysical Research Institute (NGRI), to map mineralized zones. The Aladahalli schist belt area, Hassan District, Karnataka, was proven to be a sulphide mineralized zone from integrated geophysical studies (Jayaram, 1982). There are no outcrops in the area, the lithology has been mapped from photogeological studies, and significant EM anomalies have been observed from airborne geophysical surveys. Subsequently, surface exploration over the anomalous zone with electrical, EM, IP, and SP methods (Figure 8.3) were deployed for identifying the targets for

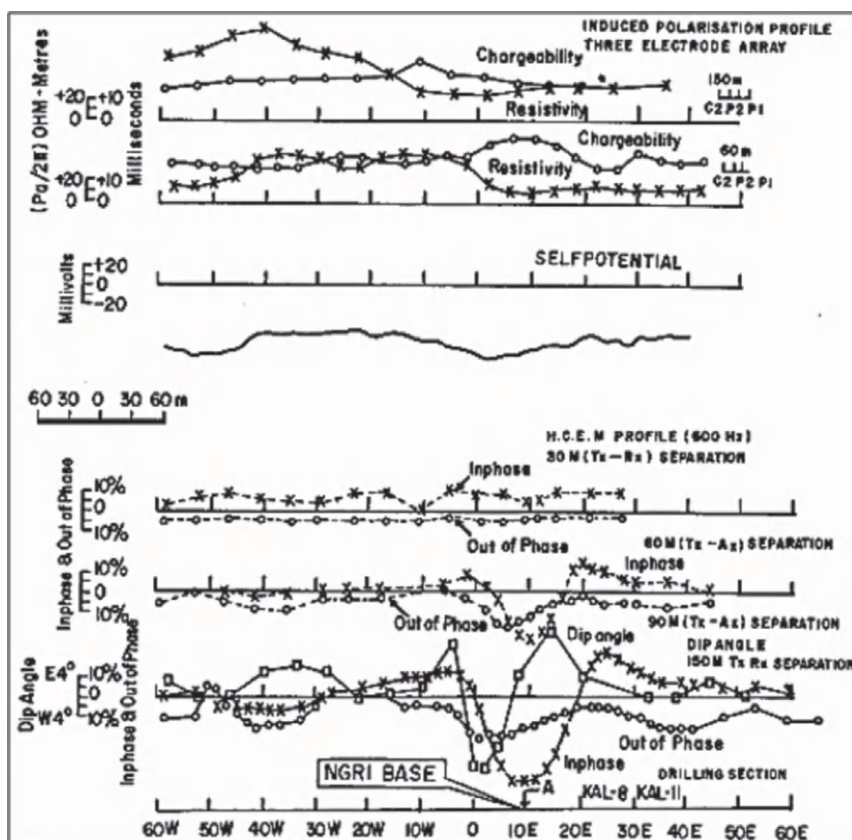


FIGURE 8.3 Various geophysical profiles—IP, SP, resistivity, EM—near the Aladahalli area, Karnataka, India. The mineralized zone (near 10E) is predicted from the integrated studies.

Source: Jayaram (1982). Courtesy of the Geological Survey of India.

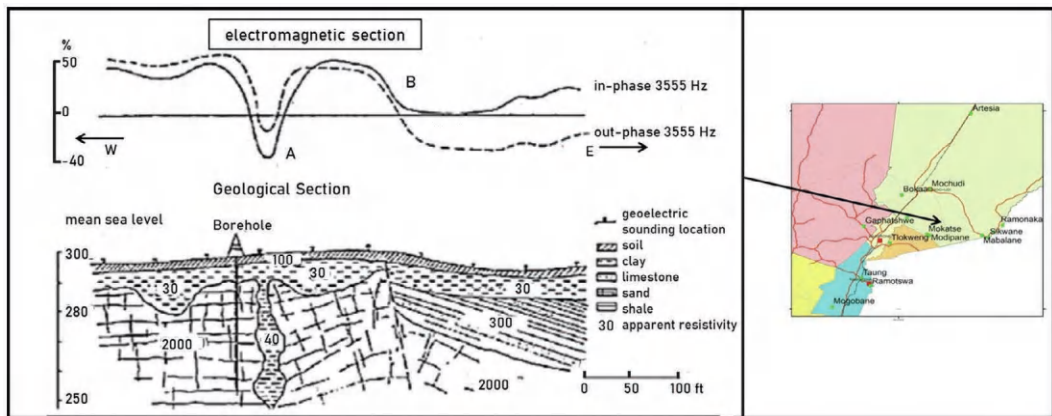


FIGURE 8.4 EMAC 36 results and geological section, Modipane, Botswana.

Source: Vogelsang (1995).

mineralization, which were validated after drilling, and about 3.9 million tonnes of ore are estimated to be in the area. Figure 8.4 gives the case histories of the electromagnetic induction method of Modipane, Botswana area (Vogelsang, 1995).

The most common use of EM induction methods is in lateral profiling, usually on traverses at right angles to the geological strike of dykes or other suspected conducting bodies. In environmental applications, it is useful for locating buried pipes that may carry fluids or gases. Recently, ground-based EM methods are used for vertical sounding, applying the same principles as in resistivity methods to obtain the conductivities of horizontal layers. Following the thumb rule, the greater the separation of transmitter and receiver, the deeper the maximum depth at which conductors may be analyzed. An important form of vertical EM sounding is the magnetotelluric method, which takes advantage of the penetrative ability of low-frequency signals from natural sources in the external geomagnetic field.

8.4.2 GROUND CONDUCTIVITY METERS (MEASUREMENTS AT LOW INDUCTION NUMBER)

It has been inferred that if the coil space is smaller than the skin depth, the term *operation at low induction numbers* is used (McNeill, 1990). Then, the in-phase component of the signal becomes very small, and the quadrature component is directly related to the conductivity of the ground. This enables a direct reading of the mean ground conductivity. However, for an interpretation of the so-obtained conductivities, we should keep in mind that the induced current flow is mainly horizontal. If electrical anisotropy due to alternate layering of good and poor conductors (e.g. clay, sand) occurs, then current flow is mainly in the good conducting layers, resulting in high measured conductivities (Seidel, 2005); that is why ground resistivity determination by electromagnetic methods can result in lower resistivities than obtained by VES measurements (Figure 8.4).

The instruments (manufactured by Geonics Ltd.) provide a direct reading of the quadrature as the apparent conductivity in mS/m. It is important to note that the secondary EM field measured in a mobile transmitter receiver system is generally a complex function of the coil spacing s , the operating frequency f , and the conductivity of the subsurface σ (Figure 8.5).

It is very important to take the ratio of the intercoil spacing s divided by the skin depth, known as the induction number B . Where the induction number is less than 1, then the ratio of the secondary to the primary of magnetic fields at the receiver is directly proportional to apparent conductivity. An

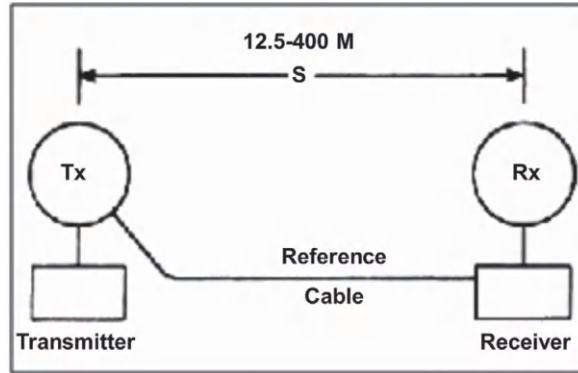


FIGURE 8.5 Moving dual-coil EM system; circles indicate the transmitter (T_x) and receiver (R_x) coils.

Source: Reynolds (1998).

important expression based on the ratio of the secondary (H_s) to primary (H_p) magnetic fields at the receiver at low induction numbers ($B \ll 1$) is given by:

$$\frac{H_s}{H_p} = \frac{i\omega\mu_0 s^2}{4} \quad (8.11)$$

The measuring system is designed to ensure that with the selected frequency f for a given inter-coil separation s , a designed response of H_p for a given transmitter is provided, the only unknown H_s , which is measured by the instrument:

$$\sigma_a = \left(\frac{4}{\mu_0 \omega s^2} \right) \left(\frac{H_s}{H_p} \right)_q \quad (8.12)$$

where the subscript q denotes the quadrature phase.

8.4.2.1 Instrument Used in the EM Induction Method

As mentioned earlier, unlike conventional resistivity techniques, no ground contact is required for EM surveys, which helps in eliminating direct electrical coupling problems and allows much more rapid data acquisition. For shallow profiling (up to 20 ft), a Geonics Inc. EM-31 terrain conductivity meter is used. One can collect as many as 10,000 data points per day with this instrument. The EM-31 uses an alternating electromagnetic field which fills the space, below- and aboveground, surrounding the transmitting coil (subsurfacesurveys.com). It is found that when the electromagnetic field couples with a conductor, for example, a steel pipe under the ground, AC eddy currents are induced to flow in the pipe, which generates a secondary magnetic field, which is sensed by the coplanar (12' offset) receiver coil. The main advantage of this methodology is that it can discriminate between the primary and secondary fields along with the outputs meant for the measurements of the secondary field associated with the phase lag and as observed by the computer on board. Therefore, a conductive zone is sensed by the induced secondary magnetic field.

An EM-34 is used for depths of investigation between 30 and 180 ft (Geophysik.uni-uenchen.de, Studium-Vorlesungen). During the survey, this instrument requires two people to operate, and up to 500 data points per day can be collected under good conditions. These tools are extremely sensitive and accurate, capable of detecting variations in conductivity of as little as 3%. Data are automatically stored in an electronic data logger for later transfer to a computer.

Among EM instrumentations, the EM-61 instrument is a high-resolution time domain device for detecting buried conductive objects. It consists of a powerful transmitter that generates a pulsed primary magnetic field when its coils are energized that induces eddy currents in nearby conductive objects. The decay of the eddy currents, following the input pulse, is measured by the coils, which in turn serve as receiver coils. The decay rate is measured for two coils, mounted concentrically, one above the other. By making the measurements at a relatively long time interval (measured in milliseconds) after the termination of the primary pulse, the response is nearly independent of the electrical conductivity of the ground, which makes the instrument supersensitive towards metal detector. Because of its unique coil arrangement, the response curve is a single having a very well-defined positive peak directly over a buried conductive object for quick and accurate location of targets. Conductive objects, to a depth of approximately 10 ft or of about 3 m, can be detected (subsurfacesurveys.com).

Data inversion to obtain the conductivity depth structure is possible if measurements were done with different frequencies or coil orientations. McNeill (1990) provided a simple method to calculate the layered ground response in which the layers below the depth Z (normalized by the coil separation) can be calculated using a sensitivity function $R(Z)$ (Figure 8.6). As an example, for a coil separation of 10 m, the depth range below 10 m ($Z = 1$) contributes to the measured apparent conductivity by 42%, whereas it is seen that the contribution of the depth range below 20 m ($Z = 2$) is only 25%. The sensitivity function $R(Z)$ can be calculated after McNeill (1990) for horizontal loops by:

$$R(Z) = \frac{1}{\sqrt{4Z^2 + 1}} \quad (8.13)$$

And for vertical loops by:

$$R(Z) = \sqrt{4Z^2 + 1} - 2Z \quad (8.14)$$

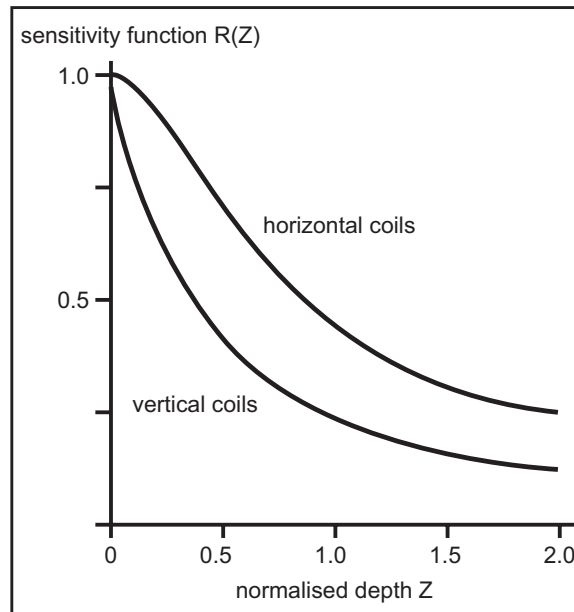


FIGURE 8.6 Sensitivity function for horizontal and vertical coplanar loops; Z = depth normalized by the coil separation.

Source: McNeill (1990).

For a layered ground consisting of n layers with conductivities σ_i and normalized depths Z_i to the bottom of layer i , the measured apparent conductivity σ_a is given by:

$$\rho_a = \sigma_1 [1 - R(Z_1)] + \sigma_2 [R(Z_2) - R(Z_1)] + \dots + \sigma_n R(Z_{n-1}) \quad (8.15)$$

The planning of field surveys, if the depth ranges of the target layers are known and a rough data interpretation is available, is possible, but the resolution of these measurements for highly resistive layers is found to be poor. EM methodology is a potential tool for mapping lateral changes of the near-surface geology with electromagnetic methods (EM-34). A sand body causes a decrease in conductivity (Figure 8.7).

Mapping deeper clay layer in the south and lower clay layer in the north with EM method (Figure 8.8).

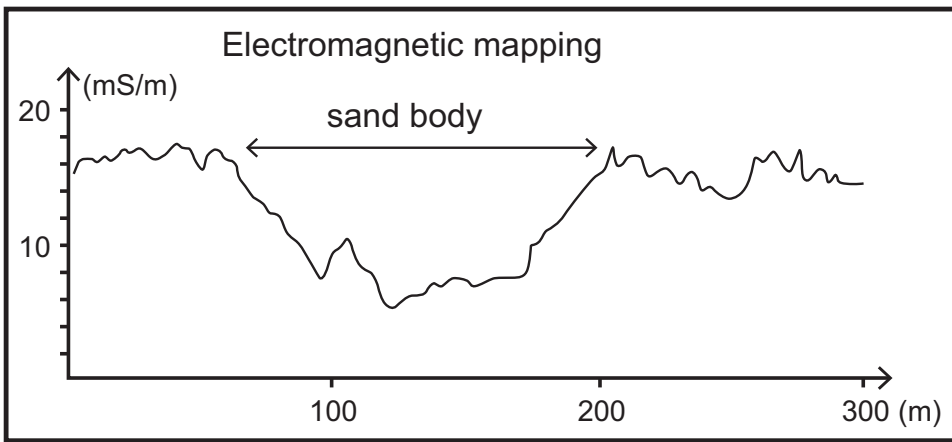


FIGURE 8.7 Lateral changes of electrical conductivity.

Source: Tezkan (1999).

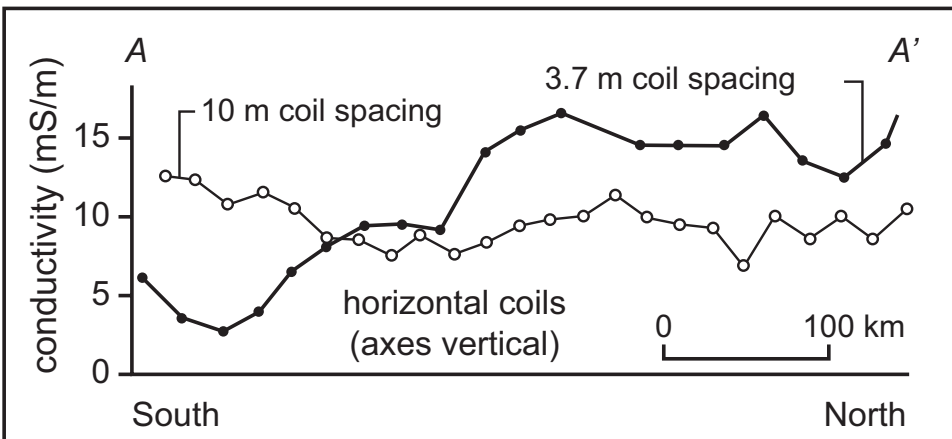


FIGURE 8.8 Clay horizon through EM method.

Source: Tezkan (1999).

Conductivity can be a good measure of changes in rock and soil units, because it is influenced by composition, porosity, degree of water saturation, and fluid chemistry. Magnetic susceptibility may provide information about erosion, paedogenesis, and anthropogenic burning, as all these processes may form or concentrate magnetic minerals (French, 2003). The use of electromagnetic induction in geoarchaeology plays an important role by estimating magnetic susceptibility by dedicated instruments, which were used in 1946 in England (Aitken, 1974). Archaeological sites can be further delineated by conducting drill holes (Dalan and Banerjee, 1998). However, such applications are often complemented by more detailed magnetic measurements (Dalan, 2007).

It is well documented, the application of EM in archaeology. This technique is now frequently used in archaeology to map large areas rapidly because of the development of instruments and software (Al-Zoubi et al., 1998; Batayneh et al., 2006; Batayneh, in press). These methods are divided into the noncontacting electromagnetic (EM) or induction methods and the soil-contacting (galvanic or resistivity) methods. Initial works with EM methods were carried out by Scollar (1962), Foster (1968), and Tite and Mullins (1969, 1970). Tabbagh (1986) has given a good summary of EM work. EM methods were also used in archaeological prospection in the United States with considerable success (Bevan, 1983; Wynn and Sherwood, 1984). The figure in Batayneh (2011) shows a survey results from EM 38 and EM 31 together with excavation at Vendel Church in Sweden (Persson, 2005).

8.5 EM METHODS USING A PLANE WAVE SOURCE OF EXCITATION

The well-known techniques belonging to this category are MT, AMT, VLF, VLF-R, RMT, and CSAMT in the far field zone. In the case of MT and AMT, electromagnetic fields of natural sources behave like plane waves at the surface of the Earth, with a small amount of energy propagating vertically downward into Earth (Vozoff, 1991). Due to their large penetration depth, these methods are normally not used in environmental geophysical exploration; only the AMT method is used for groundwater prospecting (Weidelt, 1997; Meju et al., 1999).

8.5.1 THE VERY-LOW-FREQUENCY (VLF) METHOD

Since the Slingram and ground conductivity meters are near-field methods, the VLF technique is a far-field method. The source of the VLF method is electromagnetic radiation generated in the low-frequency band of 15–25 kHz, with transmitter powers in the range of 100–1,000 kW used in long-range communication and navigational systems. The transmitted electromagnetic wave consists of electric field components E_x and E_z and a magnetic component H_y (Figure 8.9). It is important to

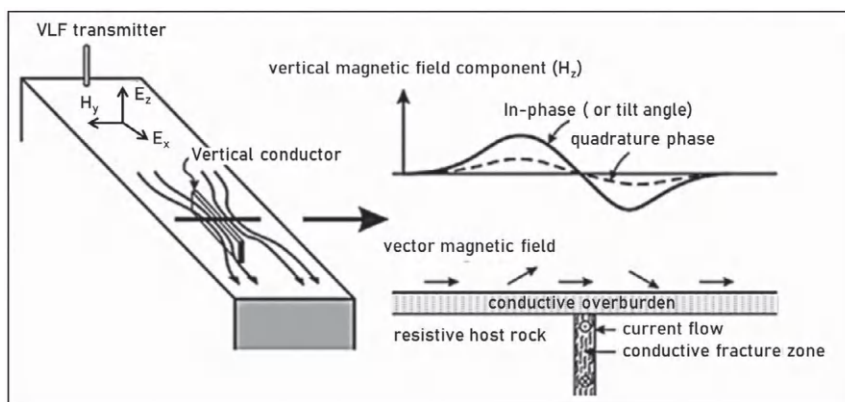


FIGURE 8.9 Left: field components from a remote VLF transmitter and electric current flow in the ground. Right: electric field component along the profile indicated in the left picture.

Source: McNeill (1990).

mention that the electric component E_x leads to current flow in the ground, especially in elongated conductive underground structures like fracture zones stretching roughly in the direction to the transmitter. These currents induce an additional magnetic field component H_z which results in a tilted magnetic field vector in the vicinity of the good conductor with a crossover at the centre of the good conductor. Usually, the tilt angle is measured by a radio receiver with the antenna tilted to the direction of maximum received signal.

Most importantly, the tilt angle data can be plotted as profiles where the crossovers correspond to narrow conducting zones. However, if the data density is sufficiently large to produce a contour map, the crossover should be converted to peak values to give a clear picture on the map. This can be done by applying an averaging filter to the data, for example, the Fraser filter, which converts a sequence of equidistant data $M_1 \dots M_4$ into

$$F_1 = (M_3 + M_4) - (M_1 - M_2) \quad (8.16)$$

F_1 is plotted at the centre of this data sequence.

There are more than a score of stations around the world transmitting VLF signals continuously for military purposes (Wilsom, 1989). Areas without shade have signal strengths, which are too weak to measure using EM method.

Figure 8.10 shows the VLF magnetic field anomalies at the margins of an extended conductor.

Figure 8.11 shows the schematic view of the data acquisition system of the VLF measurement system.

Figure 8.12 describes VLF data with out-of-phase component on a profile for groundwater exploration in Niger. The frequency used was 18.3 kHz. Borehole locations are also marked in the diagram for easy comprehension.

It is reported that the Wajrakarur kimberlite pipe 6 in Eastern Dharwar Craton is hardly explored using the latest ground-based geophysical techniques, then the study used the very-low-frequency electromagnetic (VLF-EM) method for understanding the aerial extension, depth, and geometry of the kimberlite pipe (Kumar et al., 2020). The VLF result delineates the possible kimberlite pipe at about 75 and 200 m profile distances, with depth range of about 15–45 m, having a relatively low resistivity value (53–135 Ω -m). A high-resistive zone has been recognized at a profile distance 150 m, with depth range of about 15–50 m, which is associated with the granite rock (250–532 Ω -m). It is observed that the results of the VLF-EM study are well validated using geological borehole data over the study area reported by the Geological Survey of India (GSI) (Kumar et al., 2020).

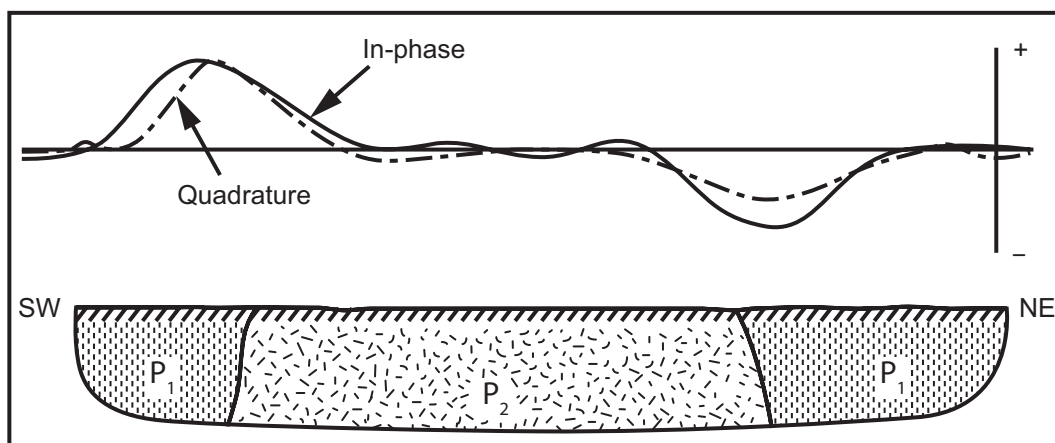


FIGURE 8.10 Anomaly over extended conductor.

Source: Wilsom (1989).



FIGURE 8.11 VLF measurements: loop antenna.

Source: Soon et al. (2015).

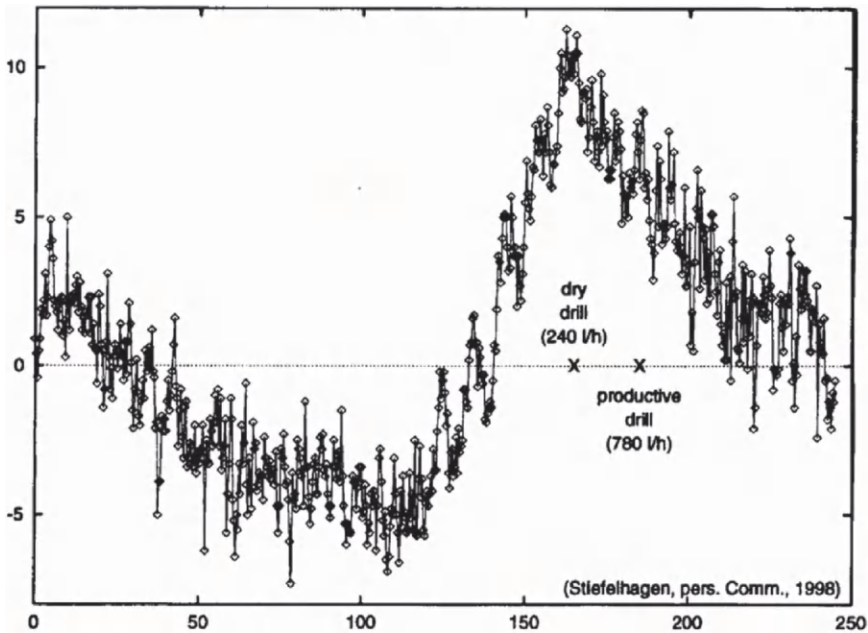


FIGURE 8.12 VLF data.

Source: Tezkan (1999).

8.5.2 VLF RESISTIVITY AND RADIO MAGNETOTELLURIC (RMT, CSRMT) METHOD

This section has been described in detail under Chapter 9. VLF, VLF-R, and RMT methods use the carrier waves from high-powered civilian and military transmitters operating in a frequency range between 10 kHz and 1 MHz (Figure 8.13). At a great distance from such transmitters (vertical electric dipole), local electromagnetic fields can be viewed as being those of a plane wave. The electromagnetic field consists of a horizontal magnetic field perpendicular to the direction of propagation and a horizontal electric field in the direction of propagation. The presence of a large discrete conductor or any anomalous conductivity structure in the Earth will modify the anomalous magnetic and electric fields observed. The ratio of the orthogonal horizontal components $\frac{E_x}{H_y}$ is related to an average resistivity of the subsurface for the observed frequency. The phase difference between E_x and H_y also contains information of the conductivity structure.

The RMT method is an extension of VLF frequencies (10–30 kHz) to higher frequencies up to 1 MHz (Muller, 1983; Lagabrielle, 1986; Turberg et al., 1994). The principle of this method is demonstrated schematically in Figure 8.13. At a large distance from the transmitters, the magnetic field is measured by a coil and compared with the electric component measured between two grounded electrodes spaced up to 20 m apart. The electrodes are placed perpendicular to the magnetic field (e.g. in the direction of the transmitter). From the ratio of amplitudes of the horizontal electric and magnetic field components, apparent resistivity values are derived using the Cagniard formula of magnetotellurics, as shown in Figure 8.13. Now:

$$\rho_a(f) = \frac{1}{\mu\omega} \left| \frac{E_x}{H_y} \right|^2 \quad (8.17)$$

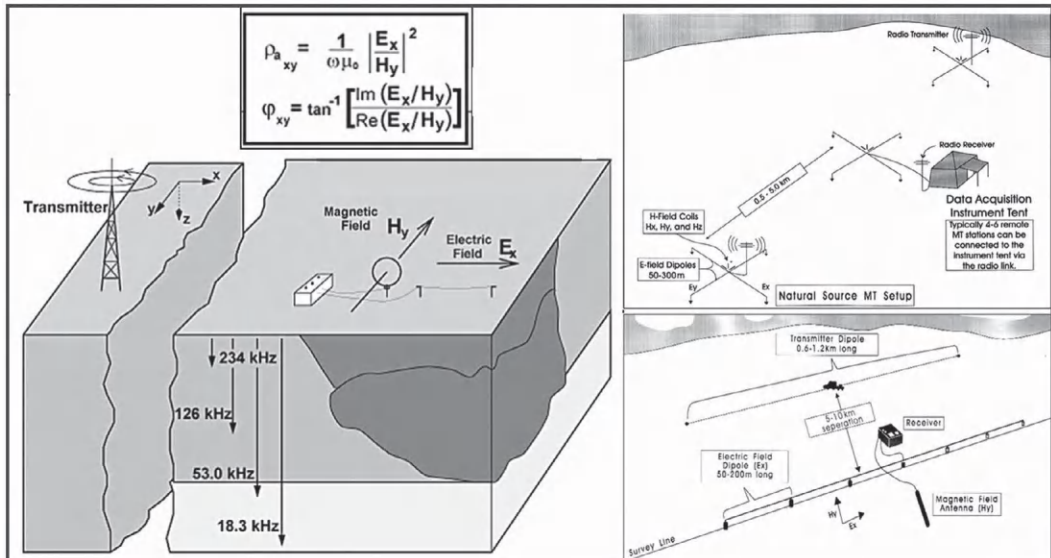


FIGURE 8.13 Schematic principle of radio-magnetotelluric sounding.

Source: Cagniard (1953).

$$\varphi(f) = \tan^{-1} \left[\frac{\text{Im} \frac{E_x}{H_y}}{\text{Re} \frac{E_x}{H_y}} \right] \quad (8.18)$$

E_x = amplitude of the horizontal electric field component, V/m
 H_x = amplitude of the horizontal magnetic field, A/m
 μ = magnetic permeability of free space, $\mu = 4\pi \times 10^{-7}$ H/m
 $\omega = 2\pi f$
 f = frequency, Hz

As the penetration depth of the radio signal depends on the frequency, a VLF-R or RMT multifrequency dataset can be inverted to the resistivity depth structure. For inversion, the apparent resistivity $\rho_a(f)$ and the phase shift $\varphi(f)$ (between E and H) are used. It is shown that the plotting of apparent resistivity and phase has close correspondence and it reflects as a function of frequency given in figure (Figure 8.14). The following information gives the interpretation of RMT phase data in tabular format (Table 8.2).

Furthermore—since the source transmitter is operating at a single well-established frequency—it is possible to measure accurately the phase angle between the electric and magnetic field components. Due to the large number of transmitters (especially in Europe), it is easily possible to cover the entire frequency range with selected frequency pairs. In an ideal case, each pair should have similar frequencies with one transmitter direction parallel (E-polarization) and one perpendicular (B-polarization) to the general strike given by geological or anthropological structures. Assuming a two-dimensional resistivity distribution in the survey area, the data can be associated to the E- and B-polarizations (Tezkan et al., 1996; Zacher et al., 1996). With this assumption, the E- and B-polarization data can be inverted jointly using 2-D inversion techniques (e.g. Madden and Mackie, 1989; Smith and Booker, 1991). Model calculations show that displacement currents can be neglected under normal conditions up to 1 MHz, and the plane wave approximation is valid for VLF and RMT data (Zacher, 1992; Schroder, 1994), so that the well-tested magnetotelluric interpretation software can be used.

In practice, however, the strike direction is often unknown. In these situations, the same procedure is applied (i.e. perpendicular transmitter pairs with similar frequencies are chosen), and the data are interpreted by 2-D conductivity models, which can be biased by 3-D effects. VLF-R data are normally presented along profiles or as contour maps of resistivity and phase. In general, no sounding information is available, and interpretation is carried out mostly qualitatively. However, in

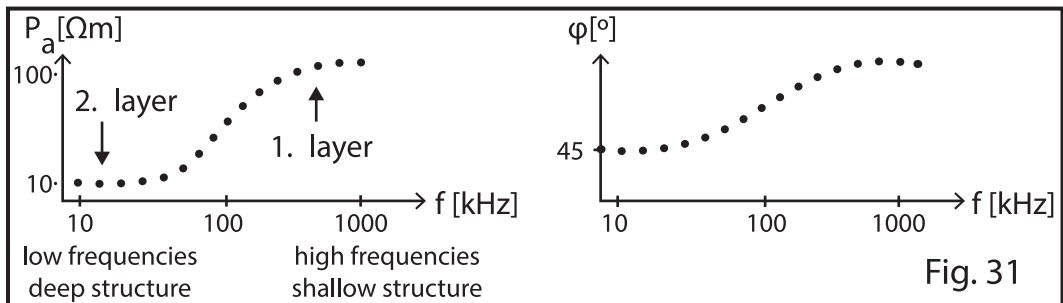


FIGURE 8.14 RMT data of ρ_a and $\varphi(f)$.

Source: Tezkan (1999).

TABLE 8.2
Interpretation of Phase Data

Phase Angle	Inferences	Remarks
$\varphi > 45^\circ$	Corresponds to both poor conductor (ρ_1) and good conductor (ρ_2) when phase angle greater than 45°	$\rho_1 > \rho_2$
$\varphi < 45^\circ$	Corresponds to both good conductor (ρ_1) and poor conductor (ρ_2) when phase angle less than 45°	$\rho_1 < \rho_2$
$\varphi = 45^\circ$	Homogeneous half-space encountered at phase angle 45°	P

some cases, the data can also be interpreted in terms of a 1-D and 2-D conductivity model (Ogilvy et al., 1997; Beamish, 1998). VLF resistivity can be considered as an alternative for terrain conductivity methods in many environmental and engineering surveys. Regular VLF (e.g. the measurement of the transfer functions for the local vertical magnetic field) is normally used to locate large mineralized zones but can also be useful in delineating lateral inhomogeneities of conductivity (e.g. shear zones and overburden troughs).

VLF and RMT instruments are lightweight and easy to use. This is one important reason for their wide use in environmental applications. Using the RMT instrument developed by the Hydrogeological Institute of the University of Neuchatel (Muller, 1983), scalar measurements (i.e. apparent resistivity and phase observations from one transmitter direction) in the frequency range between 1 kHz and 300 kHz can be carried out in a relatively short time (about 2 minutes are necessary to measure the apparent resistivity and phase values for four frequencies at one station) (Bosch and Gurk, 2000). Hollier-Larousse et al. (1994) have also developed an RMT system (10 kHz–1 MHz), but only apparent resistivities (no phase information) for selected frequencies are measured. Several impressive applications of their technique on engineering problems have been reported (e.g. Dupis and Choquier, 1996). There is no instrument available which is suitable for carrying out tensor measurements for frequencies above 70 kHz. The Stratagem system of Geonics allows tensor measurements up to 70 kHz. VLF and RMT systems can only carry out scalar measurements. The development of an instrument operating in a broad frequency range from several Hz to 100 MHz will be a task for the future.

In general, RMT surveys are carried out on undisturbed geology and on buried waste deposits. Therefore, a priori information about the overall resistivity of the host medium is usually available for RMT applications. The 2-D inversion of an RMT field dataset observed on a waste site near Cologne, Germany, by using different inversion algorithms, confirms some of the results derived by model studies using synthetic data. The following example shows a survey result of the RMT method in which the lower boundary was derived from drillings of a waste site near Cologne (Tezkan, 1999). The figure in Tezkan (1999; Candansayar and Tezkan, 2006) shows the apparent resistivity data on profile 95°N for three frequencies at the Cologne waste site. The lateral boundaries at profile meter 30 and 130 coincide with the drilling results. The transmitters are located in EW direction. The figure in Tezkan, 1999; Candansayar and Tezkan, 2006) gives 2-D inversion result for profile 95°N at the Cologne waste site which is characterized by low resistivities ($<30 \Omega\text{-m}$) between profile 30 and 120. The deeper low resistive structure indicates the clay layer. The figure in Naskar et al., 2012) shows a 2-D MT crustal model (TE mode) along Borgaon–Barwah, north of the Central Indian Shield (CIS). Apart from the conductive feature, there are a couple of vertical fault-like features near stations 4 and 9. The Lithosphere–Asthenosphere (LA) boundary is interpreted to be around 90–110 km depth (Naskar et al., 2012).

8.6 TRANSIENT ELECTROMAGNETIC (TEM/TDEM) TECHNIQUES

With the advent of new technology on frequency modulation during the 1950s–1960s, a controlled source frequency domain electromagnetic (FDEM) sounding became widely used in exploration. In erst years, to construct a sounding curve with frequency variations, we used a monochromatic transmitted field, as well as keeping changes the separation between transmitter and receiver in the history of electrical geophysics. The method is detailed in various works, including Frischknecht (1967), Vanyan et al. (1967a, 1967b), Keller (1968), Wait (1982), and Kaufman and Keller (1983).

Beginning in the 1960s, however, interest developed in the use of EM sounding methods to explore the depths of importance for oil and gas reservoirs and for geothermal systems. It has been shown that the penetration of the EM field is governed by the extent of electric resistivity of the subsurface materials that can pierce even beyond several kilometres for the subsurface media associated with comparatively lower resistivity. Considering that the depth of investigation scales by the inverse wave number, $\frac{1}{k} = \frac{1}{(i\omega\mu_0\sigma)^{\frac{1}{2}}}$, the frequency content transmitted by the controlled source

decreases by the square of the depth to be investigated. Moreover, the frequency must be lowered in proportion to the square of the conductivity, inasmuch as typical oil and gas or geothermal exploration problems involve Earth resistivities that are lower by factors of ten or more than the average resistivity of the Earth. The range of using frequencies from a few hertz to a few kilohertz has now been extended to a broadband frequency range from a fraction of hertz to a megahertz, which helps study new problems.

It is observed that generating a large-amplitude alternating current is not an easy task at megahertz frequencies. It is also found that the sequential observation of field components at various frequencies in the megahertz range can require an excessive amount of time. In order to overcome these problems in geophysical exploration, obvious solutions to both practical problems are found in using a time domain signal. For example, a signal generated by a step-on/step-off current in the transmitter contains a wide spectrum of frequencies within a short transmitting time.

To obtain a greater depth of investigation was a challenge since it was in transition from frequency domain (FD) to time domain (TD) EM methods. It is observed that time domain EM (TDEM) has advantages over FDEM with respect to a greater depth of investigation while the source and receiver are close together. This fact seems to be quite obvious today; however, it took some time and effort for geophysicists to understand the physics and mathematics of the phenomenon. It has been demonstrated that FDEM has limitations in deployment for the near zone of transmitters due to a lack of sensitivity towards the detection of electrical conductivity of the subsurface media. It was observed earlier that the transition from the frequency domain to the time domain is as a simple superposition of the frequency domain signals within a wide frequency range, and one may not use the TDEM method in the near zone of the transmitter. Russian geophysicist Gregory Grigorievich Obukhov, in the late 1960s, removed the misconception by demonstrating the fact that the TDEM field in the near-zone field is linearly proportional to the resistivity of the medium (Obukhov, 1968). Unfortunately, Obukhov's original result was not well received by his peers. At that time, many prominent Russian geophysicists could not believe that the time domain pulse, which was treated as a simple superposition of the frequency domain signals, would possess such radically different properties. Nonetheless, the mathematics behind Obukhov's discovery was solid, and his result was later confirmed theoretically and practically in several field experiments. Obukhov's result can be understood as described under Appendix 1 of Chapter 8).

There was a paradigm shift in the geophysical exploration using EM methods after the spectacular study made by Obukhov (1968). In order to have local EM sounding, we generally place the receiver in close proximity to the transmitter. The depth of investigation was controlled by the time of the recorded signal in the transmitter. With the improved accuracy of the late time measurements and increased power of the source, the depth of investigation was increased as well. A practical

application of TDEM in oil and gas exploration has very proven track record. Several Russian researchers demonstrated the key roles in the development of the method of TDEM soundings in the near-source zone (Sidorov and Tickshaev, 1969; Kaufman and Morozova, 1970; Kaufman, 1989).

Thus, the time domain electromagnetics (TDEM) is a geophysical exploration technique in which electric and magnetic fields are induced by transient pulses of electric current, and the subsequent decay response is measured. TEM/TDEM methods are generally able to determine subsurface electrical properties but are also sensitive to subsurface magnetic properties in applications like unexploded ordnance (UXO) detection and characterization. It is a proven fact that TEM/TDEM surveys are a very common surface EM techniques, and these find their applications in mineral exploration, in groundwater exploration, and for environmental mapping, having applicability to resolve issues related to both onshore and offshore geophysical exploration subsurface layers of varying physical property contrasts. Two fundamental electromagnetic principles, namely, Faraday's law of induction and Lenz's law, are the backbone of TEM surveys, in which a loop of wire is generally energized by a direct current. During the processes, at some time (t_0), the current is cut off as quickly as possible; in addition, transients are induced in the conductive bodies as well. The field system for TEM/TDEM consists of a transmitter instrument, a transmitting coil or transmitting wire, a receiver coil or antenna, and a receiver instrument that gets deployment in the electromagnetic survey using a TDEM sounding.

TDEM surveys are sensitive to estimate the depth of a layer with a low electrical resistivity, for example, a clay layer or salt water-saturated sediments. The depth of penetration can be up to 500 m (Kearey et al., 2002). It is reported that TEM/TDEM measurements allow geophysical exploration from a few metres below the surface to several hundred metres of depth, depending on subsurface resistivity, current induced, receiver sensitivity, and transmitter-receiver geometry. Low-power TEM/TDEM instruments (terra-TEM from Monex Geosope in Australia) can operate using C-cell batteries, and midrange systems with a power of 2.5 kW can operate with automotive batteries. Moreover, a powerful system (20 kW ~ 150 kW) with a separate generator, usually mounted on a truck, plane, or helicopter to provide the necessary current for deep investigations, is also deployed for the TEM/TDEM.

In coastal regions, this technique is very attractive, as TDEM can potentially delineate the location of the freshwater-saltwater interface, which is of major interest for supporting freshwater resources management in coastal aquifers. Some of the potential applications of electromagnetic surveys are found for the mapping of buried channels (Auken et al., 2008) and the mapping of a cave system in Mexico (Supper et al., 2009).

It is proven fact that DC cannot penetrate through resistive formations, which is the main limitation of the resistivity method. Electromagnetic induction methods, based on transient electromagnetic fields, overcome this difficulty because a transient field can easily propagate through resistors like a radio wave propagates in the air. At the same time, a transient electromagnetic field provides information not only about the resistivity of rocks but also about two other important electromagnetic parameters: magnetic permeability, μ , and the dielectric constant, ϵ . Many different transient electromagnetic survey methods have been introduced for geophysical exploration. Among those, some of the important methods have specific applicability for specific purposes, such as the short-offset transient electromagnetic (SOTEM) and University of Toronto EM (UTEM) for exploration of ore bodies, whilst the long-offset transient electromagnetic (LOTEM) and the multitransient electromagnetic (MTEM) for oil and gas exploration.

The UTEM (the University of Toronto EM) system uses a large fixed horizontal transmitter loop as a source. An array of receivers located both inside and outside the transmitter loop measures all three components of the magnetic field (H_x , H_y , and H_z) and the horizontal components of the electric field (E_x and E_y). It is observed that the UTEM transmitter sends a low-frequency current of a precise triangular waveform through the transmitter loop (Lamontagne et al., 1978). Since the magnetic field sensed at the receiver coil is the time derivative of the transmitted magnetic field, in 'free space', a precise square wave voltage would be induced in the receiver. In the presence of subsurface

conductors, the received waveform is substantially distorted from a square wave. This distortion is used to determine the conductivity distribution in underground formations. The UTEM system has a relatively limited depth range. It is primarily applied in mineral exploration (Zhdanov, 2018).

The long-offset transient electromagnetic (LOTEM) technique was designed for deep penetration in the ground. The LOTEM system consists of a grounded wire transmitter and an induction loop and electric field receivers. It is reported that a typical distance between transmitter and receiver is approximately equal to or greater than the exploration depth. A detailed description of the LOTEM method can be found in Strack (1992). The data acquisition and processing are similar to that of a seismic system. After prestack processing, the data are stacked and finally converted into apparent resistivities after further poststack processing. The apparent resistivities are then either inverted or directly transformed into subsurface resistivity images.

Recently, demands on EM methods have been growing in the exploration for hydrocarbons (oil and gas) both on land and in marine environments. Due to the difference in resistivities between water-bearing layers (low-resistivity) and hydrocarbon-bearing layers (high-resistivity), EM methods are considerably useful tools to reduce the risk of drilling dry wells. However, compared to seismic methods, which are the primary methods of choice, the resolution of conventional EM methods was not satisfactory enough to be used practically in the exploration for hydrocarbons. After several years of research and improvement of the instruments of EM methods, a new modification to TDEM methods was introduced by Wright et al. (2002), Hobbs et al. (2005), and Ziolkowski et al. (2007); the technique was called the multitransient electromagnetic (MTEM) method.

The survey configurations of the MTEM method are very similar to those of seismic surveys with multiple receivers and multiple transmitters. The transmitter is a current bipole source, and the receivers are also a line of bipoles with two electrodes. The key characteristic of the MTEM survey is the use of multiple transmitter–receiver separations, which results in the combination of parametric sounding and profiling. The current in bipole transmitters has a waveform of a finite-length signal formed by a pseudorandom binary sequence (PRBS) of the step functions. The time-varying voltage signals with the flow of current for each of the given pair of receiver electrodes are measured simultaneously. Data are recorded with a fine sampling rate (about 0.05 ms) to improve the resolution of the MTEM survey. The equivalent impulse response (electric field) of the subsurface of the Earth is recovered from these two measurements by deconvolution (Wright et al., 2002).

The rapid interpretation of the MTEM data is based on the apparent resistivity calculation and 1-D inversion of the data obtained for every transmitter–receiver pairs. The more efficient interpretation of impulse response obtained by the MTEM method can be performed by time domain electromagnetic migration to create images of geoelectrical structure in the subsurface of the Earth, which can be used for detecting the location of mineral deposits. Finally, a comprehensive interpretation of the MTEM data can be achieved by rigorous 3-D inversion, which provides additional capabilities for successful exploration for mineral resources. Volcanogenic massive sulphide (VMS) deposits are often found on land, due to obduction processes. Geologically, VMS deposits contain Au, Ag, Cu, Zn, and Pb resources, causing distinct geophysical properties of VMS with higher density and higher conductivity than silicate rocks (Franklin et al., 2005). It was therefore why EM was one of the best-suited tools that explored VMS deposits. However, large quantities of sulphides are easily detectable by modern EM methods using TEM (Safipour et al., 2018). It is worth mentioning that continuous mineralization interconnected veinlets should be considered carefully, as the EM signal may be weaker in veinlet-style mineralization. VMS deposits are characterized by less-intense EM signals when compared to those with massive sulphides.

Skarn deposits can have economic concentrations of Fe, W, Cu, Pb, Zn, Mo, Au, Ag, and Sn, which are detectable by EM method due to the resistivity contrasts from their surrounding rocks. The Gongpengzi Cu-Zn-W skarn deposit located in Binxian County, Heilongjiang Province, NE China, was also detected by EM method. There are abundant skarn deposits in China. For instance, the Xianglushan mine, in the Jiangxi Province, represents one of the largest tungsten mines in

China in terms of its annual production of about 3,600 t, and it is considered a classic tungsten skarn deposit (Zhang and Ni, 2018). The Shizhuyuan mine, which is one of the largest tungsten deposits in the world, has also been explored using TEM/TDEM techniques (Lu et al., 2003; Zhu et al., 2015). Additionally, the TEM technique was successfully employed to discover the Sharang porphyry molybdenum (Mo) deposits in Tibet (Xue et al., 2012). Recently, the buried Precambrian banded iron formation (BIF) located at the North China Craton (NCC) was discovered using the short-offset transient electromagnetic method (SOTEM) (Chen et al., 2017). This survey area is covered by thick Quaternary sediments. Marine TEM and controlled-source electromagnetic (CSEM) methods have also been applied in the exploration of submarine metal deposits. Moreover, it is observed from earlier exploration history that the exploration of the submarine mineral deposits is not so easy and economical (Tao et al., 2013). The use of marine EM methods has, in recent years, focused on exploring submarine hydrothermal polymetallic sulphide deposits. It is reported that the EM response of submarine hydrothermal sulphide deposits is detectable by modern offshore TEM methods. Hydrothermal sulphide deposits at the mid-Atlantic midocean ridge and the south-west Indian Ocean are also reported by Li et al. (2016). Guo et al. (2020) modelled these deposits in 3-D using TEM forward modelling. Recently, the Ministry of Earth Sciences, India, has also discovered a hydrothermal polymetallic sulphide zone in the Indian Ocean under its deep ocean mission using a different technology.

8.6.1 MINERAL EXPLORATION

There are several difficulties encountered during the exploration of metal-bearing rocks. Metal ore bodies are relatively small but valuable, and many non-economic minerals can masquerade as conductive ore bodies. That is why it is very important to select a specific set of EM methods to solve a specified exploration problem. Here, we outline several of the most promising electrical and EM techniques for mineral exploration.

The time domain sounding and profiling are the primary techniques used in exploring metallic ore deposits. Time domain sounding was developed in Russia in the 1960s and, since then, has spawned a multitude of TDEM methods routinely used by exploration companies. As discussed earlier, the time domain sounding system has proven to be one of the most effective in exploration, but it is more efficient to measure the spatial time structure of the entire EM field excited by a fixed or moving transmitter, an approach called time domain array sounding (TDAS). Detailed information about the subsurface ore bodies can be obtained by synthesizing adequate EM holograms in which different locations of transmitters and receivers in various directions are used to focus the targeted ore bodies through intense illumination, which allows better confidentiality of detecting ore bodies with less ambiguity and better reliability through the solution of the inverse problem.

Time domain array sounding (TDAS) is the method of choice in mineral exploration used by several agencies, such as the Mount Isa Mines (now Geophysical Resource Service) MIMDAS system and BHP Billiton's Geoforret. Transmitters may be constructed from the horizontal loops and grounded dipoles, whereas the receivers can be horizontal loops and/or induction coils for measuring the different components of the magnetic field. The grounded dipoles are used for electric field observations. TDAS enables the same survey configuration to be used as that for TDEM, induced polarization, and MT methods. We can observe a greater amount of detailed information about the subsurface structures because of the greater number of different components of the EM field.

The controlled-source magnetotelluric (CSMT) method is one of the efficient EM methods used in mineral exploration. Simultaneous observations of electric and magnetic fields are made in CSMT, and the impedance is estimated the same as that made in the conventional MT method. Interestingly, MT sounding differs from that of CSMT in the sense that MT has the observed EM field, which is the natural MT field, whereas in CSMT we register the controlled-source EM field. The main advantages of CSMT are connected with the possibility of taking into account the near-surface inhomogeneities and inverting the observed data for 3-D geoelectrical structures. There are

some examples of successful applications of the MT method in mineral exploration (Morrison and Nichols, 1997). It is usually suggested to adopt a combination of the natural MT method for low frequencies and the CSMT method for high frequencies in conducting EM soundings so that target bodies can be mapped from the depth of the first metres down to several kilometres.

Another important application of the EM method is in cross-borehole (cross-hole) and surface-to-borehole studies. It is observed that the setting up of the transmitters and/or the receivers into the hole significantly increases the resolution of EM methods because we can illuminate the target from different directions. Several different systems are used for borehole investigations. In recent years, a very specialized technique of mapping has been developed in which a cross-hole survey under the frequency domain is one of those methodologies that work using a vertical magnetic dipole system. The transmitter is located in one of the boreholes, and the receivers are in another borehole, which can be used for locating and characterizing massive sulphide deposits or continuity of coal seams. Airborne electromagnetic (AEM) methods are used in mineral exploration (Fountain, 1998; Witherly, 2000).

It is observed that the interpretation of AEM data is still based on simple conductivity–depth transforms (Macnae et al., 1998; Fullagar and Reid, 2001), layered Earth inversions (Chen and Raiche, 1998; Farquharson et al., 2003), or laterally constrained layered Earth inversions (Auken et al., 2005). Even with widespread use, it has been repeatedly demonstrated that these inherently 1-D methods are often invalid for recovering simple 2-D and 3-D targets (Ellis, 1998), let alone anything resembling geologic complexity (Raiche et al., 2001). Applications of the 3-D parameterized inversion methods of AEM have been found very powerful in cases of thin sheets embedded in conductive hosts for certain types of targets (Wolfgram and Golden, 2001).

Wilson et al. (2010) showed how a 3-D inversion of data acquired in the field was used for salinity mapping over the Bookpurnong irrigation district in South Australia, as the area, with declining vegetation, had decreased flooding frequency, permanent weir pool levels, and recent drought. It is reported that the RESOLVE frequency domain helicopter system with six operating frequencies was flown in the area in August 2008. The transmitter–receiver separation was 7.91 m for the five horizontal coplanar coil sets and 8.99 m for the single vertical coaxial coil set. The design of survey is also very sophisticated, where the surveys were conducted approximately at 45,000 stations, distributed over 26 lines oriented in a north-west-to-south-east direction with 100 m line spacing and seven tie lines.

The processing and modelling based on 3-D inversion method were carried out by Wilson et al. (2010) when they inverted each station for a layered Earth model using the AirBeo code of Raiche et al. (2007). The figure in Zhdanov (2010) shows a slice of the model at 4 m depth derived from an interpolation of the layered Earth inversion results. The figure in Zhdanov (2010) shows the same image with the same colour scale but obtained from the 3-D inversion results. The Murray River, which has a lower conductivity than the floodplains, is clearly visible in the 3-D inversion results. The important limitation of this inverted set of data for the layered Earth model showed smearing as artefacts in some areas that completely misses the fact that the Murray River is present. This case study shows that 3-D inversion results are in far better agreement with the known geology of the area than results obtained from layered Earth inversion.

TEM surveying using the **CSIRO Transient Electromagnetic** (SIROTEM) equipment is commonly applied in nickel sulphide explorations in the Kambalda District, Western Australia (Trench and Williams, 2019). It was developed by the Commonwealth Scientific and Industrial Research Organisation (CSIRO) in Australia. Survey geometries vary considerably depending on the respective aims and size of the study area, but in general, the coincident-loop geometry is commonly deployed with loops of 50 to 200 m of length. Station spacing is generally less than the loop size. The successful application of surface EM was documented by Pridmore et al. (1984) for both the Carnilya Hill and Edwin deposits (Trench and Williams, 2019). There are difficulties in EM interpretation over structurally disrupted sulphide bodies and in areas of thick, conductive cover (e.g. salt lakes). Discrimination of sedimentary from sulphide conductors remains difficult purely from the

magnitude and decay characteristics of the EM response but may be achieved when combined with other factors (e.g. conductor geometry, geological setting, and surface geochemistry). An example of a surface TEM anomaly in the vicinity of a nickel gossan is shown in Trench and Williams (2019). It is reported that the close spatial occurrence of the gossan with sulphidic shales made it very difficult to determine the cause of the TEM anomaly from the surface data alone.

8.6.2 HYDROCARBON EXPLORATION

Several sets of studies demonstrated that the seismic method is a very traditional as well as a foremost method of choice for hydrocarbon exploration wherein electrical methods found limited application in such studies. However, even long ago in the former Soviet Union, now Russia, electrical methods played an important role in the discovery of oil and gas fields (Spies, 1980). It is observed that electrical methods are very much an effective methodology for locating hydrocarbon reservoirs in western Siberia since it was very difficult to locate by conventional reflection seismic methods. Different lithologic facies of the reservoirs cannot be detected precisely because of little structural reflection. An acoustic impedance-based seismic method of exploration helps detect the plane structures of the potentially oil-bearing subsurface strata or layers, whilst the electrical methods detect and delineate the reservoirs directly that contain oils/gas. Because of this, the integrated approach based on a joint application of seismic exploration and electrical exploration yielded a successful discovery of major Siberian oil and gas deposits (Zhdanov and Keller, 1994).

It also found a similar situation in the oil and gas provinces of the Precaspian depression in Russia. Most importantly, the oil-containing layers are located in a depth range of 3–5 km below the salt dome structures, complicating the interpretation of seismic due to ambiguous dataset. The salt layers have detectability because of very high resistivity and are transparent for the transient EM fields, which allows a joint interpretation of the seismic and EM data that produces the robust result for exploration. The EM methods applied in this case are TDEM sounding and profiling. As with mineral exploration, time domain array sounding (TDAS) seems to be the most effective tool in hydrocarbon exploration as well. Intriguingly, TDAS can be designed as a complete EM analogue, as the seismic data acquisition system consisted of multiple transmitters and receivers. It is worth mentioning that the resolution of data is dictated by the number of receivers in a network; it is therefore during survey when it is always advisable to design the network of receivers as dense as possible so that data with necessary resolution can be acquired from the densest network of receivers for the given time domain array sounding (TDAS) of the EM method. Creating such a system is a major goal for developing EM exploration methods in the very near future.

Most ground-controlled source (inductive) and natural source EM methods are strongly biased towards the more conductive parts of the geoelectric section. It is reported that the LOTEM method uses a grounded wire transmitter and generates additional vertical induction current flow in the subsurface (Strack et al., 1989). The use of electric field sensors means that information about the resistive parts of the section can be obtained. LOTEM can be successfully applied in the exploration of hydrocarbon reservoirs, where the task is to define variations in the resistive part of the geoelectric section. In the first case history of Europe, the resistivity of a resistive layer was resolved in an electromagnetically noisy environment. The case history of Australia also showed that a good seismic reflection data may help fix some of the parameters in the geoelectric model. Then joint inversion of the magnetic and electric field components allowed obtaining a porosity section of the subsurface. When more experience is available, the method could be improved further using multichannel acquisition and processing systems, as is done in seismic reflection work.

The electric field resolves the layer parameters (resistivities and thicknesses) of the top two layers, but not quite as well as the magnetic field importance (Strack et al., 1989) of between 50 and 80%. However, both the thickness of the resistive third layer and its resistivity were better resolved by the electric field data, with importance between 10 and 30% as compared to less than 1% for the magnetic field data. This difference in sensitivity to the resistive layer parameters was analyzed

directly with the Jacobian sensitivity matrix (Raiche et al., 2001, 2007). It is also reported that the blocked electric well log may serve as a representative model for the survey area to generate synthetic electric and magnetic field data. Each column of this matrix is a measure of the sensitivity of the dataset to variations in the corresponding Earth parameter. The figure in Strack et al. (1989) shows the sensitivities to variations of the parameters of the intermediate (resistive) layer. Most important characteristics of the electrical responses of different subsurface layers are noted, and it has been observed that the change of resistivity in the resistive layers has the magnetic field appearing almost insensitive to these changes of resistivity, whilst the subsurface layers associated with changes in the resistivity and thicknesses exhibit these existing strong impacts on the electric fields. Recent findings revealed that LOTEM measurements can be made successfully in the electrically noisy environment of industrialized Western Europe. The resistive layer parameters could be resolved over a producing oilfield.

In evaluating LOTEM in Australia, the capabilities of the method for resolving variations in resistive layers in the Canning Basin were of particular interest. Minor production of hydrocarbons is obtained from porous zones within the very extensive Windjana/Nullara limestone of Devonian–Famennian age. The limestone is easily mapped by seismic reflection surveys. The major exploration difficulty has been in locating zones of adequate porosity. For evaluation purposes, a region was selected where high-quality seismic sections and good well logs were available. West Kora was being drilled in mid-1985 when the LOTEM survey was undertaken. Strack et al. (1989) showed interpreted resistivity variations based on the magnetic component data only. The values shown for the resistive carbonate were poorly resolved by the inversion when only magnetic data were used. However, when the electric field data were added for joint inversion, the results shown by Strack et al. (1989) are more depictive. The resistivity of the carbonates is much better resolved. The resolution is so much distinctive that the resistive unit and porosity changes of $\pm 5\%$ would be detected. The assumption was made that water resistivity and water saturation were constant within each unit; the expected variation of resistivity with sand-to-shale ratio for each clastic unit has been calculated (Schlumberger, 1987). Using similar calibration curves for each unit allowed transforming resistivity variations to percent sandstone within clastic units. At the same time, for the carbonate, it could be assumed that Archie's law (Archie, 1942) relates resistivity variations to porosity variations (Schlumberger, 1987), what has been explained using the concept of Strack et al. (1989), which demonstrated that the results of transforming the resistivity section to sandstone values in percentage are self-explanatory. The variations in resistivity within units are seen to be small, indicating a uniformly low porosity.

8.6.3 STUDY OF A SALTWATER INTRUSION WITH TIME DOMAIN ELECTROMAGNETIC (TDEM) DATA

Different electromagnetic (EM) methods found their applications at Monterey Bay, California, where resources of water are managed by evaluating the efficacy of geophysical data (Herckenrath, 2012). The field layout of TDEM survey is shown in the figure in Herckenrath (2012). The other site is located 30 km south and is called Fort Ord. For the Monterey, California, site, 14 TDEM soundings were available, and a cross-sectional model with uniform aquifer properties is assumed (Herckenrath, 2012). Duque et al. (2008), Langevin et al. (2003), and Guerin et al. (2001) indicated that calibration may be made through saltwater intrusion models with different geophysical data, but all these approaches have been using a sequential hydrogeophysical inversion (SHI) approach (with the exception of Bauer-Gottwein et al., 2010). The geology and hydrologic processes that are represented by the saltwater intrusion model represent a conceptual hydrogeological cross section (Herckenrath, 2012) which only represents the water table aquifer that is found in this area separated from the deep aquifer system by a clay unit marked by the Lower Aromas sand formation.

The current extent of the freshwater–saltwater interface is the result of pumping activities at this site over a time period of 67 years. For the site, no water level or salt concentration data were

available, and exact properties of the present and past water supply wells are unknown. For this aquifer, Herckenrath (2012) wants to estimate five uniform aquifer properties (diffusion, dispersion, hydraulic conductivity, anisotropy, and porosity) and one petrophysical shape parameter (m in Archie's law) by fitting the collected TDEM data (more than 300 apparent resistivities). Herckenrath (2012) discussed the inverted three-layer resistivity models for TDEM sounding 1–14 as a function of the distance with respect to the coast. All electrical resistivity models show a first layer with a high resistivity, a second layer with very low electrical resistivity, and a third layer with a higher resistivity compared to the second layer. The first TDEM layer can be interpreted as a layer comprising both the dry deposits and the freshwater-saturated aquifer, where the second layer with a very low electrical resistivity of less than $1 \Omega\text{m}$ represents the saltwater-saturated sediments.

The figure in Herckenrath (2012) showed the inversion result for the 25-layer smooth model. The pattern is consistent with the figure in Herckenrath (2012), showing both the clay layer and the saltwater-saturated layer. It is observed that the methodology provided more information about the distribution of fresh and salt water in the aquifer even if in the absence of the estimate of the dip of the clay layer (Herckenrath, 2012). The figure in Herckenrath (2012) shows the resulting TDEM model using a CHI-S. The bottom layer is the same in both (Herckenrath, 2012), representing the clay layer that is present at the site. Herckenrath (2012) fixed the particular geophysical parameters for this clay layer when performing the CHI-S, as the saltwater intrusion model does not provide any information about this layer. The second commonality between the figure in Herckenrath (2012) and the geophysical inversion results is the high electrical resistivity of the top layer. The difference, however, is the much higher amount of detail for electrical resistivity in the aquifer. The electrical resistivity model resulting from the 25-layer smooth inversion has a similar resolution but shows a much less consistent pattern about the distribution of salt and fresh water in the aquifer.

8.6.4 GEOPHYSICAL IMAGING OF ALPINE ROCK GLACIERS: MURTÈL CASE STUDY

Applications of the EM method have a proven track record for the permafrost that underlies mostly the polar and mountainous regions of the Earth, like the Murtèl rock glacier (approximately $150 \times 300 \text{ m}^2$), located in the Upper Engadine area of the eastern Swiss Alps at an altitude range of 2,600–2,800 m (Maurer and Hauck, 2017). It is moving horizontally a few centimetres per year (Kaab and Weber, 2004). The ongoing movements have generated numerous flow lobes (Maurer and Hauck, 2017). In 1987, a borehole was drilled to a 58 m depth (Vonder Muhll and Holub, 1992). It is reported for the existence of large boulders and air-filled voids at 0–3 m, an ice-rich layer (more than 40% ice) at 3–15 m, a mixture of ice and sand (30–35% ice) at 15–30 m, and a mixture of sands and boulders at 30–52 m and the bedrock below. Maurer and Hauck (2017) recorded TDEM data for determining the internal structure of the Murtèl rock glacier.

A specialized transmitter loop of varying diameters from a few metres to a few tens of metres is required for the typical TDEM systems. Deploying large loops across blocky rock glaciers can be quite laborious, and only a few measurements per day may be possible. TDEM systems are best suited for soundings under these conditions because only a single layout needs to be set up as per the given sounding. Since the secondary magnetic fields decay rapidly once the electric currents are turned off, TDEM recording systems need to densely sample the fields at early times.

TDEM data provide first-order estimates of rock glacier thickness (Hauck et al., 2001). In principle, this type of information could be obtained directly from the borehole logs, but to investigate the suitability of the TDEM method on rock glaciers, a sounding was performed near the borehole (Hauck et al., 2001). Acquisition time was about half an hour. These studies used a commercial software package developed by Interpex Limited (1990) to process data acquired by TDEM.

Results of the TDEM survey are very much depictive in the figure by Maurer and Hauck (2017). Maurer and Hauck (2017) showed the observed data plotted as apparent resistivity vs time since current turn-off. Maurer and Hauck (2017) interpreted data using the general shape of the curve, and they inferred the presence of a relatively resistive layer sandwiched between two more conductive

layers, one of which approaches the surface. The solid line in the figure in Maurer and Hauck (2017) represents the best-fitting model obtained from the inversion program that explains the data nearly as well as the best-fitting model (Nabighian and Macnae, 1987). It was also reported that the significant decrease of resistivity at 54 ± 15 m depth coincides with the rock–glacier–bedrock interface. Maurer and Hauck (2017) demonstrated that the resolution capabilities of TDEM soundings in the uppermost 10 m are limited, but a careful interpretation has been suggested for the deeper layers (Nabighian and Macnae, 1987). Nevertheless, it is interesting to note that the near-surface 10 k Ω -m values shown by Maurer and Hauck (2017) are consistent with interpretation. The interpretation is constrained by information provided by the borehole. This active layer, which corresponds to loose boulders at the top of the borehole, has low seismic speeds of $<2,000$ ms⁻¹ (Maurer and Hauck, 2017).

8.7 GROUND-PENETRATING OR PROBING RADAR (GPR)

GPR is a technique of imaging the subsurface at a high resolution of data acquired using radars of antennas of different frequencies. Since the mid-1980s, GPR has become enormously popular, particularly with the engineering and archaeological communities. It was applied since the 1960s in connection with the development of radio echo sounding of polar ice sheets. For regional and large-scale investigations, radar measurements have been made from aircraft and satellites. Ancient river drainage systems buried beneath desert sands in Africa as an important source of potable water have also been deciphered using GPR techniques with greater confidence.

Recent studies suggested that the physical principle of GPR, an electromagnetic pulse reflection method, is very similar to that of the seismic reflection surveys. In the GPR method, a short radar pulse in the frequency band 10 MHz to 1 GHz is impinged into the ground. In this GPR, the EM wave reflects from the subsurface EM reflector and is recorded by a receiver for the given electrical properties of the media into which EM wave impinges (Kirsch, 2006). The physical principle of seismic wave (acoustic wave depends on the product of velocity and density of the media) and light wave (an electromagnetic wave) works on a similar principle, where reflections and diffractions happen, indicating that the electromagnetic waves occur at boundaries between rock strata due to the electrical property contrasts or differences in electrical properties. Reynolds (1998) showed that the direct and reflected amplitudes of the electric field strength E are recorded as a function of travel time. The electric permittivity (dielectric constant) μ and the electric conductivity σ are petrophysical parameters which determine the reflectivity of the layered boundaries and depths of EM wave penetration. A high pulse rate enables quasicontinuous measurements by pulling the antennas along a profile with a progress of several kilometres per hour. In many cases, a preliminary interpretation of the profile data (radargram) is possible in the field. As radio waves travel at high speeds (in air, 300,000 km/s), the travel time is a few tens to several thousand nanoseconds ($1 \text{ ns} = 10^{-9} \text{ s}$). This requires a very accurate instrumentation to measure the signal. The antennae used can be in monostatic or bistatic mode.

Monostatic. One antenna device is used as both transmitter and receiver.

Bistatic. Two separate antennae are used, with one serving as the transmitter and the other as the receiver.

8.7.1 ELECTRIC PERMITTIVITY AND CONDUCTIVITY

Electromagnetic wave propagation is governed by Maxwell's equations. These four coupled differential equations, which have already been discussed previously as well as in Chapter 9, provide the interrelationship between the electric field E and the magnetic field H , in time-, space-, and

material-related equations, such as Ohm's law. It has been shown that the relevant material properties for electromagnetic wave propagation are the electric permittivity μ and the specific electric conductivity σ . To be consistent with the usual notation in electrodynamics (von Hippel, 1954a), we introduce a quantity called complex permittivity ϵ^* , which takes into account both the loss-free displacement of electric charges in its real part and the dissipation (heating) caused by dielectric and conduction losses in its imaginary part:

$$\epsilon^* = \epsilon' - i\epsilon'' \quad (8.19)$$

with $i = \sqrt{-1}$

Normalization with the vacuum permittivity $\epsilon_0 = 8.8542 \times 10^{-12} [\text{AsV}^{-1}\text{m}^{-1}]$ yields the dimensionless relative complex permittivity:

$$\epsilon_r^* = \frac{\epsilon^*}{\epsilon_0} = \epsilon_r' - i\epsilon_r'' \quad (8.20)$$

Note that the real part ϵ_r' is equal to the commonly used permittivity. Here, we will use ϵ as an abbreviation for ϵ_r' . The conductivity σ is related to ϵ_r'' and the frequency f by

$$\epsilon_r'' = \frac{\sigma}{\omega\epsilon_0} \quad (8.21)$$

with $\omega = 2\pi f [\text{s}^{-1}]$.

We can define the loss tangent, which is nothing but the ratio of conduction to displacement currents, as

$$\tan \delta = \frac{\epsilon_r'}{\epsilon_r''} = \frac{\sigma}{2\pi f \epsilon_0 \epsilon_r} \quad (8.22)$$

For $\tan \delta = 1$, the conduction currents are equal to the displacement currents. Electromagnetic wave propagation utilized in GPR takes place for $\tan \delta < 1$, while for $\tan \delta > 1$, diffusive processes govern (the case of low-frequency electromagnetics like RMT or AEM). Typical values for electric permittivity and conductivity of different aquifer materials are given in the following (Table 8.3).

Examples for $\tan \delta$ of three materials may illustrate the use of eq. 8.22. The properties of dry sand ($\sigma = 1 \text{ mS/m}$, $\epsilon = 4$), saturated sand ($\sigma = 10 \text{ mS/m}$, $\epsilon = 25$), and wet clay ($\sigma = 50 \text{ mS/m}$, $\epsilon = 25$) are inserted at a frequency of 50 MHz. For dry sand, we get $\tan \delta = 0.09$, which makes it

TABLE 8.3

Physical Properties of Aquifers and Impermeable Layers (*)

	Seismic V_p , m/s	Geoelectric, Electromagnetic		GPR (*)	
		Resistivity (Ωm)	Conductivity (mS/m)	Permittivity	Wave Velocity (cm/ns) (mean)
Gravel, sand (dry)	300–800	500–2,000	0.5–2	3–5	15
Gravel, sand (saturated)	1,500–2,000	60–200	5–17	20–30	6
Fractured rock	1,500–3,000	60–2,000	0.5–17	20–30	6
Solid rock	>3,000	>2,000	<0.5	4–6	13
Till	1,500–2,200	30–60	17–34	5–40	6
Clay	1,500–2,500	10–30	34–100	5–40	6

Source: after Davis and Annan (1989).

well suited for GPR. The saturated sand with $\tan\delta = 0.14$ is still fairly good, but the clay shows $\tan\delta = 0.72$, which is at the limit for wave propagation. The magnetic permeability μ of geological materials has little effect with respect to GPR applications. The permeability of free space can be expressed as:

$$\mu = \mu_0 = 4\pi \times 10^{-7} [\text{VsA}^{-1}\text{m}^{-1}]$$

8.7.1.1 Complex Permittivity of Water

It is reported that the bulk conductivity of geological materials is a high-frequency effect due to the dipole character of water molecules (which is used for heating in the microwave oven). This phenomenon is called Debye relaxation. Some important studies, namely, by von Hippel (1954b) and Hoekstra and Delaney (1974), define the real part of the permittivity ε'_r of water drops from about 80 at low frequencies down to 5.8 at 1,000 GHz; the imaginary part goes through a maximum at about 10 GHz, which corresponds to a high-frequency conductivity of more than 20 S/m. Although the frequencies involved in GPR for hydrogeological mapping do not exceed 400 MHz, Debye relaxation occurs in moist or saturated materials already at 100 MHz. That is why low frequencies provide deep penetration.

8.7.2 ELECTROMAGNETIC WAVE PROPAGATION

Important and simple solutions for Maxwell's equations are harmonic plane waves in a rectangular coordinate system, the electric field $\mathbf{E} = (E_x, 0, 0)$, and the magnetic field $\mathbf{H} = (0, H_y, 0)$, with propagation in z direction (e.g. von Hippel, 1954a):

$$E_x(z, t) = E_{x0} e^{\gamma z - \omega t} [\text{Vm}^{-1}] \quad (8.23)$$

$$H_y(z, t) = H_{y0} e^{\gamma z - \omega t} [\text{Vm}^{-1}] \quad (8.24)$$

With the complex propagation constant:

$$\gamma = \alpha + i\beta [\text{m}^{-1}] \quad (8.25)$$

The E-field, then, is written as:

$$\mathbf{E} = E_0 e^{i(\beta z - \omega t)} e^{-\alpha z} \quad (8.26)$$

In the previous equations, E and H are orthogonal to each other and orthogonal to the direction of propagation. The first exponential function contains the phase variations of the wave with respect to time and space. Hence, β is called the phase constant. The second exponential term describes an exponential attenuation of the wave with the attenuation constant α . It is observed that both parameters are frequency-dependent and are related to the complex permittivity via

$$\alpha = \frac{\omega}{c_0} \sqrt{\frac{\varepsilon'_r}{2} \left(\sqrt{1 + \tan^2 \delta} - 1 \right)} [\text{Neper m}^{-1}] \quad (8.27)$$

where *attenuation*, α , nepers or dB/m, is the term applied to the reduction in signal.

$$\beta = \frac{\omega}{c_0} \sqrt{\frac{\varepsilon_r'}{2} \left(\sqrt{1 + \tan^2 \delta} + 1 \right)} \quad [\text{rad m}^{-1}] \quad (8.28)$$

with the speed of light in vacuum (or air) $c_0 = 2.998 \times 10^8 \text{ m/s} = 300 \text{ m}/\mu\text{s} = 0.3 \text{ m/ns}$. Instead of the attenuation factor α , the absorption coefficient $\alpha' = 8.686 \alpha$ (in dB/m) is used. Note that α is small if $\tan \delta \ll 1$, which means that the conductivity σ has to be small for low absorption.

The unit dB (decibel) is a logarithmic measure for ratios of output power or output voltage to input power or input voltage (1dB = 20 log, that is, output amplification/input amplification), which can also be expressed in terms of intensity (I), as intensity is proportional to the square of amplitude, as 1dB = 10 log (output intensity/input intensity). In the case of amplitude of voltage, 20 dB means a ratio of 10, 40 dB means a ratio of 100, and so on. In the case of power as intensity, 10 dB means a ratio of 10, 20 dB means a ratio of 100, and so on. It is observed that dB has the advantage of avoiding huge numbers if the ratios get large. The propagation velocity v (phase velocity) of the radar waves is calculated from eq. 8.28 by

$$v = \frac{2\pi f}{\beta} \quad [\text{m ns}^{-1} \text{ for } f \text{ in GHz}] \quad (8.29)$$

which collapses to

$$v \approx \frac{c_0}{\sqrt{\varepsilon}} \quad (8.30)$$

for low-loss materials with $\tan \delta \ll 1$.

The wavelength λ is calculated from the well-known relation $v = \lambda f$ for low-loss materials as

$$\lambda = \frac{c_0}{f \sqrt{\varepsilon}} \quad (8.31)$$

For a low-loss ground with $\varepsilon = 9$, we calculate from eq. 8.30 that the velocity in the ground is one-third of the speed of light: $v = 0.1 \text{ m/ns}$. It can be exemplified that at a frequency of $f = 100 \text{ MHz} = 0.1 \text{ GHz}$, the wavelength in the ground will be 1 m. To give examples for the frequency dependence of the propagation parameters absorption α' and velocity v , we again look at the properties of dry sand, saturated sand, and wet clay as at the end of Section 8.7.1.1. In the figure in Reynolds (1997; Annan, 2009), v and α' are plotted versus frequency in a double-logarithmic scale. Reynolds (1997) and Annan (2009) showed that the calculation of absorption can be done with and without the effect of Debye relaxation. This approach highlights the contrast between real-world behavior and a simplified perspective that considers only DC conductivity.

It is found that velocity and absorption of the electromagnetic waves in the low-frequency region rise with the square of frequency. For higher frequencies, the velocities reach constant values, and it can be shown that this takes place for $\tan \delta < 1$. To get dispersionless propagation, one should take care that the GPR frequency is high enough to meet the condition $\tan \delta = 0.5$. If there was no water in the materials, the absorption would show plateaus for higher frequencies. However, the Debye relaxation starts to raise absorption significantly above 100 MHz, indicating the fact that the high-frequency pulses propagate in a wet or less-dry conditions or environment, which is subjected to severe attenuation of the EM wave. From the figure in Kirsch (2006; von Hippel, 1954a), it can be learned that frequencies around 50 MHz provide a good trade-off with respect to absorption and dispersion (frequency dependence of propagation parameters).

8.7.3 REFLECTION AND REFRACTION OF PLANE WAVES

At a plane boundary between two media 1 and 2 with different properties of ε and σ , an incident electromagnetic wave is both reflected in 1 and refracted into 2. The waves have different directions with respect to their incidence, reflection, and refraction, and these all fall in one plane in such a way that the angle of incidence θ_i becomes equal to the angle of reflection θ_r' . The angle of refraction θ_r is related to the angle of incidence θ_i by Snell's law. The index of refraction n_{21} can be expressed in relation with the permittivity of the medium:

$$\frac{\sin\theta_i}{\sin\theta_r} = n_{21} = \sqrt{\frac{\varepsilon_2}{\varepsilon_1}} = \frac{v_1}{v_2} \quad (8.32)$$

It infers the refraction from lower to higher velocity that can be understood when the wave passes from the subsurface with velocity V into air with velocity C_0 , then there is a critical angle θ_c where θ_r reaches 90° , which is nothing but a case of total internal reflection in which the refracted wave grazes to the surface by propagating along the interface of the two strata having physical property contrasts in terms of velocity. For all $\theta_i > \theta_c$, there is total internal reflection. In the far field of a transmitting antenna, electromagnetic waves are approximated by plane waves. We generally consider only differences of ε in the special case of normal incidence ($\theta_i = 0^\circ$), which is contained in the more general Fresnel's equations (von Hippel, 1954a). Thus, the coefficient of reflection or reflectivity (reflected part of incident unit amplitude) r is given by

$$r = \frac{\sqrt{\varepsilon_1} - \sqrt{\varepsilon_2}}{\sqrt{\varepsilon_1} + \sqrt{\varepsilon_2}} = \frac{v_2 - v_1}{v_2 + v_1} \quad (8.33)$$

and the refracted (transmitted) part t by

$$t = \frac{2v_2}{v_2 + v_1} = \frac{2\sqrt{\varepsilon_1}}{\sqrt{\varepsilon_1} + \sqrt{\varepsilon_2}} \quad (8.34)$$

where ε_1 and ε_2 are the relative permittivities (dielectric constants) of the two media separated by the interface, and v_1 and v_2 the radar velocities within them. It is observed from Table 8.4 that the velocities of geological materials fall within the range 0.06–0.160 m/ns for a typical value for $f = 100$ MHz.

If conductive (lossy) materials are involved, even the simple case of normal incidence, the reflection coefficient may become complex, leading to a deformation of the reflected wave. Following are the material parameters given in tabular form (Table 8.4).

Water has a dielectric constant of 81. The water content of materials exerts a strong influence on the propagation of a radar pulse. Velocity contrast between dry and wet sand is useful for ground-water exploration. The velocities of geological materials are very high, which indicate corresponding wavelengths are in the range of decimetre and very good resolving of small structures. Kirsch (2006) showed that the wavelength of electromagnetic waves was independent of signal frequency. Multiple reflections are found bouncing between the surface and a strong reflector, which are not significant for GPR because there is usually considerable absorption in the soil and rock.

8.7.4 SCATTERING AND DIFFRACTION

The process of scattering and diffraction in GPR is similar to that of the reflection seismology, where diffraction occurs at discontinuities of reflectors (interrupts, faults with a shift) when dimensions of objects would be smaller compared to the wavelength. Low frequencies reduce the disturbing influence of inhomogeneities in the rock (geological noise) if the wavelength resulting from the

TABLE 8.4
Material Parameters

Material	Relative Permittivity, ϵ_r	Conductivity, σ (mS/m)	Velocity (m/ns [†])	Wavelength (m)	
				At 50 Hz	At 1,000 Hz
Air	11	0	0.30	6	0.3
Water, fresh	81	0.5	0.033	0.66	0.033
Water, sea	81	3,000	0.01	0.2	0.01
Ice, pure	3.2	0.01	0.16	3.2	0.16
Clay, wet	25–40	50–100	0.5–0.6	10–12	0.5–0.6
Granite	4–6	0.01	0.1–0.12	2–2.4	0.1–0.12
Limestone	4–8	0.5–2	0.1–0.12	2–2.4	0.1–0.12
Sand, dry	3–6	0.01	0.15	3	0.15
Sand, wet	20–30	0.1–1	0.06	1.2	0.06
Shale	5–15	1–100	0.09	1.8	0.09
Silt	5–30	1–100	0.07	1.4	0.07

Source: [www. Geophysik.uni-muenchen.de](http://www.Geophysik.uni-muenchen.de) -> Studium -> Vorlesungen.

* Because radar waves take such a short time to reach the shallow interfaces, velocities are given in this form (1 ns = 10⁻⁹ s).

† All values, which are from various sources, are approximate.

equation $\lambda = \frac{c_0}{f\sqrt{\epsilon}}$ is considerably larger than the size of the inhomogeneities. It is required to have a compromise with respect to resolution. In order to localize linear diffractors as, for example, the dispositions of pipelines or cables, the concept of scattering and diffraction of EM wave has been used at high-frequency ranges. The polarization of the electric field should be parallel to the target objects for maximal sensitivity. Once the orientation of the targets is not known, then the measurements must be made along orthogonal profiles.

8.7.5 HORIZONTAL AND VERTICAL RESOLUTION

It is observed that the resolution is a measure of the ability to distinguish between signals from closely spaced targets. In ground-penetrating radar, the resolution depends on the central frequency (or wavelength, which is proportional to the pulse period) and on the range of frequencies (bandwidth) of the target (size, shape, and orientation). Important also are the coupling to the ground, the radiation patterns of the antennas (especially the diameter of the Fresnel zone), and the noise conditions in the field. As a rule of thumb, the vertical resolution is theoretically one-quarter of the wavelength $\lambda = \frac{v}{f}$, where v is the velocity of the electromagnetic wave in the medium and f is the frequency. Because the velocity is lowered by the presence of water, the resolution will be better in wet materials. The actual resolution is in the order of one-half of the wavelength. Care must be taken that the objects can be separated laterally if their distance is larger than the diameter d_f of the first Fresnel zone. For example, a GPR system operating at $f = 200$ MHz in an environment with $v = 0.1$ m/ns has a wavelength of $\lambda = 0.5$ m. The horizontal resolution at a depth $h = 4$ m will be $d_f = \sqrt{2\lambda h} = 2$ m, which is much poorer than the vertical resolution of better than 0.5 m. Horizontal resolution can be improved by data processing with migration or SAR (synthetic aperture radar, the same process as migration) to a theoretical value of $\frac{\lambda}{4}$ in the case of real 3-D measurements.

8.7.6 RAY GEOMETRY AND TRAVEL TIMES

The propagation of electromagnetic waves can be described by a ray representation as in optics and reflection seismic. As, for example, a simple horizontal two-layered model with a reflector at depth, h requires four wave paths and travel time curves. The ray path scheme for GPR is shown in the figure in Kirsch (2006; Burger et al., 2023). It is observed for two direct waves with different phase velocities and amplitudes traveling along the ground surface, the airwave and the ground wave (Banos, 1966). Then it is a known fact that the airwave travels with the largest possible velocity for electromagnetic waves—the speed of light in vacuum c_0 ; it can be used to determine time zero (like the time break in seismic surveys). The velocity v in the uppermost stratum is determined from the ground wave. Thus, changes in the ground wave indicate changes in the uppermost stratum (e.g. moisture content, type of rock). Based on the concept of Banos (1966), the travel time of the airwave t_a and the ground wave t_g as a function of distance x can be expressed as:

$$t_a = \frac{x}{c_0} \text{ and } t_g = \frac{x}{v} \quad (8.35)$$

Because v is always less than c_0 , a lateral wave is generated when the critical angle $\varphi_c = \arcsin\left(\frac{v}{c_0}\right)$ is reached, analogous to the head wave in refraction seismic surveys. As discussed earlier, then this wave propagates in air parallel to the ground surface. The critical angle φ_c is related to the critical distance x_c , which is given by

$$x_c = \frac{2hv}{\sqrt{c_0^2 - v^2}} \quad (8.36)$$

Under conditions, in such a way that for $x > x_c$, the lateral wave exists with the travel time t_l :

$$t_l = \frac{x}{c_0} + \sqrt{\frac{1}{v^2} - \frac{1}{c_0^2}} \quad (8.37)$$

It is proven fact that the refracted waves originating at subsurface boundaries are rarely observed, because in most cases, the velocity decreases with depth. The most important wave is the reflection at the layer boundary at depth h (Figure 8.15). It is similar to the seismic surveys, where the travel time t_r is given by the hyperbola:

$$\left(\frac{L}{2}\right)^2 = \left(\frac{x}{2}\right)^2 + h^2$$

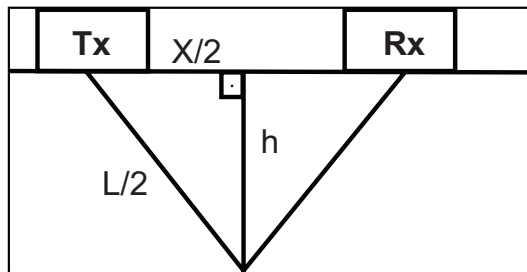


FIGURE 8.15 Diagram used in calculation for reflection.

$$\frac{L^2}{4} = \frac{x^2}{4} + h^2$$

$$L = \sqrt{x^2 + 4h^2} \text{ which implies}$$

$$t_r = \frac{1}{v} \sqrt{x^2 + 4h^2} \quad (8.38)$$

If we put a condition for $x = 0$, then we can get the zero offset travel time $t_0 = \frac{2h}{v}$. The travel times are calculated in nanosecond (ns) if the velocities are in m/ns and the distances are in m. A travel time diagram of the different types of waves is shown in the figure in Kirsch (2006; Reynolds, 1998).

8.7.7 AMPLITUDES/ATTENUATION

For the interpretation of GPR measurements and for the estimation of penetration depth, it is useful to know the main features of amplitude decay with distance from the source.

The source emits a spherical wave into the lower half-space. The amplitude decreases inversely proportional with distance due to geometric spreading and is subject to absorption due to the attenuation coefficient $\alpha = \frac{\alpha'}{8.686}$, which is dependent upon the electric conductivity σ , magnetic permeability μ , and dielectric permittivity ε properties of the geological medium as well as the frequency of the radar signal. Hence, a wave reflected at depth h with a reflectivity r and has travelled a distance $2h$ and its amplitude E_r compared to the amplitude at unit depth E_l will be

$$\frac{E_r}{E_l} = \frac{r}{2h} e^{-2\alpha h} \quad (8.39)$$

It is found that for small values of h the effect of geometric spreading is dominant, while the exponential function is stronger at larger depths. Smaller values of α for air, ice, and granite provide an ideal condition for deploying GPR.

Notably, the field strength of the direct wave diminishes in the far field with the square of the distance from the source (Banos, 1966), whilst the ground wave is decreased in addition by absorption and scattering. Moreover, Brekovskikh (1980) showed that the lateral wave decreases in the far field region approximately with the square of the distance.

It is reported for the diffractors that the so-called radar equation (e.g. Daniels, 1996) is a suitable alternative to eq. 8.39 for waves diffracted back towards the source:

$$\frac{P_r}{P_t} = q_s \frac{G^2 \lambda^2}{(4\pi)^3 h^4} e^{-4\alpha h} \quad (8.40)$$

where P_t is transmitted power (W), P_r is received power (W), q_s is scatter cross section (m²), G is antenna gain (relative to that of a spherical dipole), λ is wavelength at the centre frequency (m), whilst h represents the distance between the antenna and the diffractor (m) and α the attenuation coefficient (m⁻¹).

8.7.8 ESTIMATION OF EXPLORATION DEPTH

To avoid dispersion, the operating frequency should be chosen according to $\tan \delta < 0.5$. Rearranging eq. 8.22 for frequency and letting $\tan \delta = 0.5$, we get

$$f_m \geq \frac{36000}{\rho \epsilon} \quad (8.41)$$

where f_m is the central frequency in MHz, ρ the DC resistivity in $\Omega\text{-m}$ (inverse of conductivity σ), and ϵ is the real part of the relative permittivity.

As per conceptual physics, the variability of attenuation coefficient with frequency is seen as low enough that it negates the effect of Debye relaxation, which can be estimated from the DC conductivity (σ) and the relative permittivity (ϵ). Solving eq. 8.27 with the approximation $\tan \delta \ll 1$ yields

$$\alpha' = 1640 \frac{\sigma}{\sqrt{\epsilon}} \quad (8.42)$$

with α' in dB/m and σ in S/m.

For a successful application of GPR, the two-way absorption along the ray path (from the transmitter down to the reflector and back to the receiver) should not exceed a maximum of 60 dB. It is also observed that other losses due to geometric spreading, reflection, and scattering sum up to occupy the remaining part of the dynamic range of existing GPR systems.

Hence, we can try to put up a rule of thumb for the maximum exploration depth $h_{\max}$ by letting $2h_{\max}$; $\alpha' < 60$ dB or, after rearranging,

$$h_{\max} \leq 0.018 \rho \sqrt{\epsilon} \quad (8.43)$$

with $h_{\max}$ in m and ρ in $\Omega\text{-m}$.

In case of dry sand, saturated sand, and wet clay, frequency found changes in a minimum range of 9, 14, 72 MHz, respectively, in accordance with eq. 8.41 earlier. Using eqs. 8.42 and 7.43, the absorption coefficient for dry sand is 0.82 dB/m, and the exploration depth 36 m; for saturated sand, the respective values are 3.3 dB/m and 9 m; and it has been estimated using the preceding equation that for wet clay, the absorption is found to be 16.4 dB/m for the depth of exploration of 1.8 m.

8.7.9 TECHNICAL ASPECTS OF GPR

8.7.9.1 Overview of System Components

A very composite set of GPR system has a pulse generator, an antenna for transmission of electromagnetic pulses, along with a second antenna for reception of the direct and reflected signals, and receiving electronics to amplify and digitize the waveforms to be recorded and displayed, as described by Wu and King (1965) and Kirsch (2006). In some cases, both antennas are contained in one housing or there is one single antenna with a fast switch for changing between transmission and reception. Due care has been taken in the sense that timing and triggering of the measurement process are carried out by a control unit, which may also interface the external components, like a survey wheel or GPS. Since the cables connecting the different parts of a system are preferably nonmetallic fibre-optic cables, if metallic cables are to be used, some ringing or internal reflections may occur that require proper shielding, for which ferrite beads are used to suppress such artefacts, though it is found that these components may look different for systems of different GPR manufacturers, for which the principles of operation and even electronic circuit are generally the same for time domain (i.e. short pulse) measurements. Recently, a novel class of frequency domain equipment having the facility of stepped frequency radar has now been manufactured that uses a different concept but yields similar outcomes.

8.7.9.2 Antennas and Antenna Characteristics

It is found that broadband antennas are needed to transmit and receive short electromagnetic pulses because of the broad frequency spectrum contained in short waveforms—for a monocycle pulse, the

bandwidth is typically in the order of the central frequency. It is further indicated that the conventional broadband systems with directional antenna gain that are used for radio and television in the VHF and UHF ranges are not suitable for short pulses.

Several case studies highlighted the fact that central frequency f_c between 20 and 1,000 MHz can be used for geological and engineering investigations. Usually, a broad bandwidth is generally achieved by resistive loading (damping) of electric dipoles or by using extended electric dipole elements. Several antenna designs, such as dipoles with hyperbolic resistive loading (Wu and King 1965), unshielded and shielded bowtie antennas, and combinations of bowtie and loop dipoles with resistivity loading (Kirsch, 2006), are being used in EM exploration. Wu and King (1965) suggested that antennas with loading have a signal loss of 20 dB per pair relative to a simple dipole. The radiated wavelet is the time derivative of the excitation function of the pulse generator, which is the shortest possible pulse. The important features of different antennas reveal that the frequency spectra of loaded loop dipoles and bowtie designs are narrower, and the pulses are slightly longer, whose waveforms are similar to the second derivative of a Gauss function (Ricker wavelet), with some ringing, and the amplitude losses per antenna pair in the range of 10 dB. There is a great challenge in how to conduct measurements in buildings and below power lines and trees when signals are affected by 'air reflections' from objects in the upper half-space. Under such conditions, shielded antennas are used to suppress these unwanted disturbances, for which shielding is accomplished by placing absorbing materials and/or metal boxes above the dipoles.

GPR antennas couple to the ground by the concentration of electric field in the lower medium; the best coupling is achieved by laying the antenna flat on the ground (Annan et al., 1975). Typical far-field antenna patterns for seven ground-penetrating radar 241 short dipoles were used by Kirsch (2006) for two orientations. Finite difference time domain (FDTD), a very efficient numerical method, is applied to estimate the depths of exploration for the GPR exploration surveys (Radzevicius et al., 2003; Lampe et al., 2003).

8.7.9.3 Electronics

8.7.9.3.1 Pulse Generators

Recently, pulse generators for producing short, high-energy impulses for use in the field are often designed with transistor switches using the semiconductor avalanche effect, which allows charging of a number of condensers in parallel, and then discharged to the antenna occurs in series by the switches. Kirsch (2006) showed that pulses with nanosecond rise times and amplitudes of up to 2 KV can be produced for deep sounding, in which amplitudes of 400 V are recorded with repetition frequencies of about 100 kHz as a typical range for most commercial systems. Antenna frequency and type dictate the pulse length of a transmitted electromagnetic wave, which is lesser than 20 ns, since the pulses are generated and repeated with high accuracy. It is worth mentioning that the timing noise (jitter) within radar systems used for geophysics is usually better than 1 ns.

8.7.9.3.2 Receiver Systems

We need to have sampling of the waveform at the receiving antenna that can be done by analogue sample and hold converters or fully digital systems. In order to keep portability of the equipment and less power consumption, most systems utilize an analogue sequential sampling method with digital control. Sequential sampling suggests that only one point of the total time series (trace) is recorded per transmitted pulse. This is not economic for the pulsar, but it is for the low power system in totality. It is also reported that continuous measurements are possible because the pulse rates are very high and the pulse generators offer a long lifetime. Since the sequentially sampled signals are in the frequency range of audio waves so that they can be digitized by a 16-bit A/D converter and stored in computer memory, it is found that the achievable dynamic range of good sampling devices is 80–90 dB, which is low compared to seismic systems. We usually set the sensitivity limit by

controlling the electronic noise due to the large bandwidth, which can only be reduced by stacking (averaging of traces). It is demonstrated that digital real-time receivers used for special purposes record one trace per shot, which has the slight advantage that high-amplitude pulse generators can be *used for higher stacking rates*.

It is observed that the time window for hydrogeological investigations ranges from several 10 ns (for travel paths of several metres) to 1,000 ns for a maximum depth of about 60 m in unsaturated or 30 m in saturated low-loss material. As mentioned elsewhere in this chapter, the sampling intervals depend on the central frequency used in the study. We have to set sampling in such a way that one cycle should be sampled at least at 10 points. Thus, for a 100 MHz pulse with 10 ns cycle time, a sampling interval of 1 ns is sufficient. It is observed that most commercial GPR systems have only one to four channels; multiple-coverage measurements with different source–receiver distances (offsets) like those in reflection seismic surveys are done sequentially. In general purposes, georadar measurements are carried out as a profiling with constant offset (single coverage). It is noteworthy that the dynamic range of GPR systems is given by some manufacturers as the performance factor *PF*, which is the logarithmic ratio of the peak voltage at the given pulse generator and the minimum recordable voltage at the corresponding receiver in dB. In the present scenario, the performance factors of 130–140 dB are typical for existing systems. The *PF* is of little use to estimate penetration depth.

8.7.9.4 Models of Data Acquisition

8.7.9.4.1 Standard Procedure

There are two main procedures used for GPR surveys: constant (common) offset (CO) profiling and common midpoint measurements (CMP), or common depth point (CDP), or common reflection point (CRP), which have replaced wide-angle reflection and refraction (WARR), which are similar to the seismic method of exploration as described by Annan and Chua (1992). The methodology illustrates these two modes, CO and CMP or CDP. In the typical mode of operation of the common offset (CO) mode, we maintain the separation between the receiver and the transmitter at a fixed distance and move along a line to produce a profile (bistatic configuration). It is suggested for the necessary antenna separation (offset) from centre to centre, which is typically about one wavelength in free space for unshielded antennas and about half a wavelength for shielded designs. It is a notable fact that, if only one antenna with transmit/receive switch is used, a true zero offset section is recorded (monostatic configuration) at the cost of penetration depth. The dead time is considered at the beginning of the time window. Generally, measurements are taken at constant intervals along the profile triggered by a survey wheel or at constant time intervals at a rate of 1 to 100 measured traces per second proceeding at several kilometres per hour. It is possible and convenient to make measurements by car or tractor, which enhances the capability of GPR surveys with its high resolution, making GPR one of the fastest methods of near-surface geophysics.

Several EM explorationists (Reynolds, 1998; Benson et al., 1983) illustrated the presentation used when a graphic recorder is used to record analogue data. The blackened portions of the record during the recording of data suggest that both negative and positive excursions are in excess to the threshold values. Such presentation is adequate for most tasks where target detection is the object and post-survey processing is not anticipated. Secondly, it is observed that wide variations in the appearance of the record are possible, depending on the gain settings used during the field surveys.

In relation to seismic shooting using common depth point (CDP) or common reflection point (CRP) or common midpoint (CMP) during surveys, the transmitter and receiver are to be moved stepwise and simultaneously away from a fixed midpoint to obtain a sequential multifold coverage. The evaluation of travel time curves from CMPs gives velocity–depth functions, which are necessary to calibrate the conversion of travel time into depth. WARR soundings (not recommended) keep the transmitter at a fixed location while the receiver is moved away from it. Other methods to get velocity information are the use of borehole data (geological information, depth to groundwater)

or, for very near-surface applications, the evaluation of diffraction hyperbolas originating at underground objects like pipes. In recent years, our target has been to avoid pitfalls in the application of GPR, for which a careful judgment is required to decide the suitability of an area or target for GPR measurements through checking by deploying a test measurement, or by resistivity data from geoelectric soundings, from which absorption can be estimated using eq. 8.42. It is a proven fact that layers with high conductivity (low resistivity) prevent successful soundings due to less penetration because of the concentration of the electrical flux on the surface. It is useful to avoid some areas associated with specific features, such as heavily fertilized fields, roads paved with slag, and soils with high clay content. Some GPR systems provide band-pass filtering of the data during acquisition for signal enhancement. However, the filter settings should be kept as broadband as possible to maintain the information content during the data acquisition stage. Modern systems have gain-setting options and a stacking function, which should be carefully set to optimize data quality.

8.7.9.4.2 *Special Techniques*

Based on the law of achieving resolution of GPR data, the 3-D GPR surveys are designed for structural imaging of the subsurface strata with the profile spacing as dense as possible (a quarter wavelength) to avoid spatial aliasing. The amount of data, the necessary processing steps, and the possibilities for representation are comparable to 3-D seismic surveys. Repeated 3-D measurements are able to detect subsurface changes with time and are often referred to as 4-D measurements. The monitoring of dense non-aqueous phase liquids (DNAPL) spills shows the potential of this technique (Brewster and Annan, 1994; Brewster et al., 1995).

It is interesting to mention that just like controlled-source borehole-to-borehole (cross-hole) or borehole-to-surface seismic imaging or tomography, the radar tomography also operates from borehole to borehole or from borehole to surface with specialized slim watertight borehole antennas to image the subsurface structures within the zone of influences cordoned by a given set of transmitters and receivers. The principles of tomography require a very dense ray coverage of the illuminated area (Mishra et al., 2003). It is reported that the theoretical limit of resolution (Williamson, 1991; Williamson and Worthington, 1993) depends upon wavelength and distance from the borehole and is in the order of the first Fresnel zone, which in turn suggests that the resolution is high closer to the boreholes and lowest in between. Some successful case studies (Hubbard et al., 1997; Troncke et al., 2002) are reported.

8.7.10 DATA INTERPRETATION

It is a proven fact that careful evaluation of site conditions for both geologic or stratigraphic tasks for the targeted specific interests is required for GPR surveys, which will help achieve the desired results. If the objectives of a survey are poorly drawn, often the results of the GPR survey may be excellent records that do not have any straightforward interpretation. However, a wise strategy is to tune a GPR system such that exceptional subsurface detail is visible on the record. The geologic evaluation problem is that, except in special circumstances (like the fore-set beds inside sand dunes), there is no ready interpretation. GPR records generated through calibrated response provide distinct insight into detailed stratigraphy embedded with a thin interbedding of alluvium or even with small moisture concentration. Following is the schematic of a set of targets surveyed by GPR.

8.7.11 RANGE OF APPLICATIONS OF GPR

At this stage of integrated geophysical exploration, GPR has found its wide-scale applicability in addressing a variety of problems associated with different media in which EM wave propagates, such as **rock, soil, ice, fresh water, pavements, and structures**. A classical study carried out under a joint Indo-Canadian research project by the Central Mine Plan and the Design Institute Limited

(Coal India) and Mining Geological and Metallurgical Institute of India (MGMI), in association with Laval University Canada, detected underground coal cavities, flushing of mines successfully using GPR with 50 MHz and 100 MHz antennas (Seguin et al., 1994). Therefore, GPR is capable of detecting subsurface objects, changes in material property contrasts, along with the presence of voids and cracks. Here are some common applications of GPR:

- Contaminant plume mapping
- Landfill investigations
- Location of buried fuel tanks and oil drums
- Detection of natural cavities and fissures
- Void detection
- Ice thickness mapping
- Delineation of buried archaeological objects

Ground-penetrating radar (GPR) is a nonintrusive method of surveying the subsurface to investigate underground utilities, such as concrete, asphalt, metals, pipes, cables, or masonry. This non-destructive method uses electromagnetic radiation in the microwave band (10 MHz to 2.6 GHz; 1 MHz = 10^6 Hz, and 1 GHz = 10^9 Hz) of the radio spectrum and detects the reflected signals from subsurface structures. Reflected energy occurs at boundaries where subsurface electrical properties change. It penetrates deeper (lower frequencies, for example, <500 MHz), but with reduced resolution. It is also useful for tasks like geological mapping. Figure 8.16 shows the GPR applications and frequencies used.

GPR profile was carried out in the mining centre of Khouribga (Morocco), and it has been reported that sterile bodies are detectable and delimited using the GPR method, which was made as shown earlier for the detection of sterile bodies in a trench that provided valid geological attributes of the area. The GPR surveys were reportedly carried out using a 40 MHz antenna moving on the ground with a 3 km/h speed and 10 cm/sample of resolution (El Assel et al., 2011).

In recent years, like that of 3-D seismic imaging called 3-D seismic tomography, the concept of 3-D GPR tomography or imaging of engineering subsurfaces has become very popular because of extracting detailed three-dimensional information through imaging of the subsurface. This

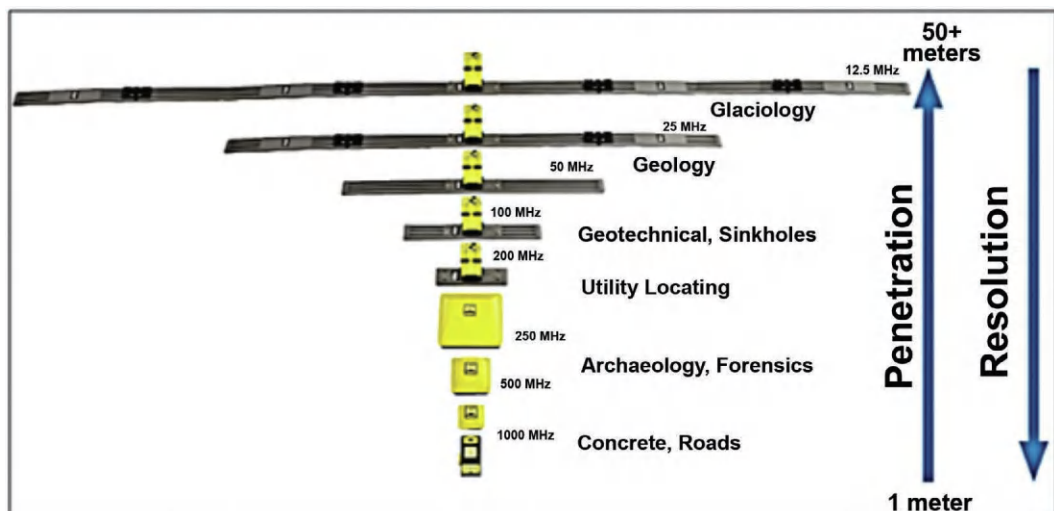


FIGURE 8.16 Utility of GPR and associated frequencies.

technique is highly used to locate archaeological structures, pipes, and cavities (Teixido et al., 2010). They demonstrate that with a 400 MHz antenna used, the field swept in parallel profiles spaced every 0.5 m, archaeological material accumulated approximately between 0–0.75 m and 0.75–1 m depth was conspicuously imaged, which explained that these accumulations of material correspond to the reflections between the archaeological structures and the filler material (Teixido et al., 2010). A GPR survey was carried out by the Geological Survey of India (GSI) using (1) 200 MHz frequency, (2) 400 MHz frequency, and (3) 80 MHz frequency (MLF antenna) (Arctic GSI, 2015). Kirsch (2006) showed that GPR survey for sand/clay boundary was clearly associated with the reflector at 12 ns.

Giovanni (2012) shows the high resolving power of the method GPR, which is allowed to highlight the irregular alternation of layers related to EM characteristics very different (corresponding to limestones and marly clays). It also identifies two main systems of folds with a large radius of curvature and with varying wavelength. Kirsch (2006) used GPR profile to map a water table reflection in Nubian Sandstone, Eastern Sahara. The GPR survey was conducted near the Barajadeuli area over a stabilized sand dune located near the beach ridge (Rout et al., 2018), where they used the postprocessed GPR image based on deconvolution method in Radon software and showed the section parallel to the elongation of the sand dune whilst it was nearly perpendicular to the beach ridge (Rout et al., 2018). It is worth mentioning that the enhanced 2-D GPR profile found excellent correlation with the subsurface lithology borehole data. In 2-D GPR profile, the top layers of the sand dune are clearly seen with alternating very light-blue-red-coloured and thin horizontal lines up to a depth of 6 m (Rout et al., 2018). At a depth of 2 m, there is a light-coloured horizontal line representing the presence of shallow groundwater level. This can also be called the ‘perched/local groundwater table’ due to its presence above the clay beds located within the sand dune layers. This is evidenced in the field in the form of slightly indurated, wet, oxidized, and dark-yellow-coloured dual sand representing the oscillation zone of this perched water table, noted while making trench and collecting sand samples after conducting the GPR survey (Rout et al., 2018). In the field of cultural heritage, GPR with high-frequency antenna is also used for investigating historical masonry structures, detecting cracks and decay patterns of columns and detachment of frescoes and historic cemeteries (Masini et al., 2010).

8.8 DEEP PROBING THROUGH GROUND-PENETRATING RADAR (GPR) BY GEOLOGICAL SURVEY OF INDIA (GSI): CASE STUDY

Detailed glaciological studies on the Hamtah glacier, Lahaul–Spiti District, Himachal Pradesh (HP), were carried out by the Geological Survey of India (GSI) for the 18th consecutive year. A study of the recessional pattern of the glaciers of the Bhaga basin, Lahaul–Spiti District, HP, was undertaken for the Panchi Nala I and Panchi Nala II glaciers. The Geological Survey of India (GSI) has been participating in expeditions to Antarctica since 1981, and currently, the research stations in the Antarctica (Maitri and Bharati) are maintained by the National Centre for Polar and Ocean Research (NCPOR), Ministry of Earth Sciences (MoES), India, besides the conduct of regular Indian scientific expeditions to Antarctica and the Arctic. Currently, NCPOR-MoES is now leading the research related to the cryosphere of both the polar and oceanic regions. The GSI team carried out thorough field studies on the shear zones of the Schirmacher Oasis, and about 200 oriented samples have been brought for further studies in the laboratory to gain deeper insight into the tectonic and metamorphic history of the region. Routine glaciological observations were carried out, revealing an accumulation of 65.5 cm of snow during 2007–2008 on the ice shelf along the Princess Astrid Coast. Overall average recession of the Dakshin Gangotri glacier snout in the Schirmacher Oasis during the same period was recorded to be 1.10 m. Two members from GSI participated in the first Indian Arctic expedition held in March 2008 (GSI Annual Report).

The Geological Survey of India (GSI) has been engaged in carrying out the most serious scientific applications of GPR instrument in the field of glaciology. Swain (2018) suggested that the GPR survey can be performed by using multiple low-frequency (MLF) antennas with a range of 16 to 80 MHz frequency in continuous mode. This method was tested in and around Schirmacher Oasis, East Antarctica, to delineate the interface between the water column and bedrock in the polar lakes and between the ice and bedrock below the polar ice sheet, and deep probing of the polar ice thickness could be investigated. In Antarctica, the GPR profile obtained by using 200 MHz shows good resolution, which can be utilized to undertake the survey for estimating the lake depths of the polar lakes (Swain, 2018). When the GPR antenna frequency is lowered, the depth of investigation increases, but the resolution gets poorer. The lower frequency range in between 16 and 80 MHz penetrates to a deeper level, but the data used are to be taken in point mode. Continuous profiling was not possible earlier. Slightly deeper penetration was possible by using 100 MHz antenna frequency in bistatic mode (Swain, 2018).

For deeper penetration, the GPR survey can be performed by using multiple low-frequency (MLF) antennas with a range of 16 to 80 MHz frequency in continuous mode (Swain and Goswami, 2014). This method can be applied by the use of a platform with fixed interchangeable positions for transmitter and receivers to be kept at a fixed separation during the survey, depending upon the depth of penetration (Swain, 2018). Interestingly, another technique involved fixing the transmitter and receiver at a fixed position at the front and rear side of the snow vehicles and survey wheel attached to the unit for keeping track of the distance, which was exclusively used during the first Indian scientific expedition to the South Pole (Swain, 2010, 2018).

Though the resolution is poor by using MLF antenna ranging from 16 to 80 MHz frequency, it is able to penetrate to a deeper level. The polar ice sheet thickness estimation survey during the GSI Field Season Program (FSP) 2016–2017 shows an ice thickness of more than 420 to 580 m (Swain, 2018). Wherever the thickness was more than 580 m, the interface between ice and bedrock could not be obtained as it has reached the optimum level in that condition. During a recent 6.4 km long survey using MLF antenna with 35 MHz frequency, antiformal and synformal features were observed in the snow layers up to 40 m (Swain, 2018). To validate the data, the area was compared with the data obtained through the Bedmap2 project. The thickness map of ice cover generated using Bedmap2 data indicates that the thickness is maximum of the order of 1,200 m to 1,700 m in the central part of the study area, whereas the northern part shows thin ice cover of ice sheet that ranges up to a maximum of 722 m (Swain, 2018). Moreover, the exact estimate of ice thickness in the polar region has become a very crucial task because of several constraints; as such, a series of integrated geophysical studies have been deployed to study the different aspects of the Arctic and Antarctic region by a galaxy of scientists.

The range of the depth estimated by the Bedmap2 project is shown by a double arrow line, and the average is by a filled square in the figure in (Swain, 2018). Both of the depth estimations are comparable with the errors ranging from 2.8 to 8%. The lower the depths, the lesser the error percentage, and vice versa (Swain, 2018). Several limitations are observed in GPR methods. Some of these limitations include its dependence upon the dielectric contrast between the mediums; the presence of englacial boulders resulting in ambiguities in the interpretation of data; the presence of englacial channels within the glacier; the presence of high-conductivity materials, such as clay soils and contaminated soils, especially for geological investigations, causing signal scattering in heterogeneous conditions; the radio signal contamination due to other signals, including transmission lines, mobile towers, etc.

8.9 DELINEATION OF LEAD (Pb) CONTAMINATION FROM GPR DATA

A research was conducted to characterize subsurface contamination caused by the leaching of lead (Pb) from batteries disposed of at the surface, spread with groundwater movement. The contaminated soils and aquifer are located in the Matano formation (Sulawesi, Indonesia) (Dinanta

et al., 2023, Elsevier 100055). Ground-penetrating radar (GPR) was used to detect and delineate Pb-contaminated soils (solid-phase Pb) (Dinanta et al., 2023), since solid and aqueous phases have different characteristics and responses when subjected to electromagnetic (EM) waves. A series of studies have so far used GPR to map solid-phase lead (Pb) contamination in the subsurface. GPR uses EM as the base medium to receive subsurface images and is useful for the detection of solid-phase Pb contamination but cannot detect aqueous-phase Pb. It is observed that the remediation process involves, first, delineation of solid-phase Pb contamination in the subsurface using GPR, which required a geotechnical survey to support GPR. Dinanta et al. (2023) collected 15 lines of GPR measurements and 11 soil samples to quantify Pb, for which frequencies of 600 MHz and 900 MHz were used for the GPR antenna. The section profiles of the GPR data were acquired for various survey lines (Dinanta et al., 2023). The results identified a clay layer at a depth of between 3 m and 4 m, and Pb contamination was mostly limited to a depth of 1 m to 2 m.

The ultimate purpose of this work was to develop a 3-D delineation of Pb contamination at the study site (Dinanta et al., 2023, Elsevier 100055); this was expected to fill the gap between subsurface mapping using GPR and the distribution of contaminant read by X-ray fluorescence (XRF). In this study, Dinanta et al. (2023) used 15 traverses to construct a 3-D profile, followed by Sandmeier (2016), Ekko_Project Team (2015), Wang et al. (2020), and Voxler Team (2015). To measure Pb concentrations with depth using X-ray fluorescence (XRF), the GPR survey clearly showed that Pb contamination had migrated both vertically and laterally, but that Pb concentrations were greatest near the source area and decreased with lateral and vertical distances. The laterite soils at the site can be described as sandy loam to silty clay, with more gravel near the surface. Subsamples of the soil boring sent for laboratory analysis of Pb concentrations were in close agreement with XRF results (Dinanta et al., 2023, Elsevier 100055). Recently, a 3-D model was generated (Dinanta et al., 2023, Elsevier 100055) by overlaying the lithological interpretations of soil samples with the 3-D GPR tomography profiles.

It is now evident that the 3-D GPR model, including lithology at the site and Pb concentration, can effectively map Pb contamination in the subsurface. It is suggested that future research investigate the use of artificial intelligence to help identify significant patterns in GPR images of the subsurface. It can now be emphasized that GPR survey combined with XRF analysis of soils samples in the field is very much apt to delineate and characterize Pb contamination in soils and aquifers to make site remediation decisions.

8.10 DEVELOPMENT OF HANDHELD STEP-FREQUENCY GPR USING FIELDFOX NETWORK ANALYZER AND ITS APPLICATIONS

NGI's (NGI Oslo, Norway) classical handheld step-frequency GPR has recently been developed for carrying out geotechnical investigations in the field. The Fieldfox-based GPR can be very much suited for nondestructive testing of roads, bridges, dams, river dykes, and testing inside buildings, tunnels, etc. Kong et al. (2012) used a handheld step-frequency GPR using the Fieldfox network, and they showed a very clear rebar image that worked in the frequency band used for this test (from 700 MHz to 3,000 MHz).

The newly developed handheld GPR has been used in several countries, including China, India, Bangladesh, and Vietnam. In China, NGI's handheld GPR was successfully used to detect the structure of the Qian Tang River dyke in Hang Zhou. The purpose of the investigation was to identify the different stone layers used in the construction of the dyke and to detect the interface between stone and soil (Kong et al., 2012). In Bangladesh and Vietnam, GPR investigations (Kong et al., 2012) were successfully performed on existing river dykes. Their purpose of study was to locate weak spots in dykes so that these could be strengthened before the onset of the next monsoon season. Kong et al. (2012) showed the system worked very well in Dhaka, Bangladesh, using 1.2 m long antennas as well as for the Red River dyke in Vietnam, where weak spots were detected in the

subsurface. It was subsequently observed that there were some sinkholes on the sides of the dyke beneath the detected weak spots.

In India, the handheld GPR was successfully used to check the thickness of the road pavement in the Central Road Research Institute (CRRRI) in New Delhi. The results from the GPR tests at CRRRI show that the bitumen surfacing thickness is about 40 mm, followed by a bituminous course of around 120 mm. The reflections in the radargram show the different pavement layers and the depth (Kong et al., 2012, Proceedings of Indian Geotechnical Conference, Delhi, Paper No. A161). The Fieldfox-based GPR can be very useful when detections with very high resolution—for example, nondestructive testing of roads and bridges; testing inside buildings, tunnels, and platforms; etc.—are needed. It is also advisable to perform tests for low-frequency bands using long antennas so that a GPR can very well be used for low-frequency applications.

8.11 GEOPHYSICAL IMAGING OF ALPINE ROCK GLACIERS BY GEORADAR SURVEY: MURTÈL CASE STUDY

It is reported that the high electrical resistivity of most mountain permafrost terrains is a positive characteristic for georadar surveying in Murtèl. In such condition, the georadar techniques are most sensitive to variations in dielectric permittivity, which govern the propagation speeds of georadar waves. The high resistivities allow antennas with dominant frequencies as low as 1 MHz to be employed (Gogineni et al., 1998). For high-resolution investigations, antennas with dominant frequencies between 20 and 100 MHz have proven to be suitable (Arcone et al., 1998; Isaksen et al., 2000; Lehmann and Green, 2000; Berthling et al., 2003; Moorman et al., 2003). It is worth mentioning that proper coupling of the antennas to the ground is an important prerequisite for recording high-quality data, which needs attention in the winter, when the snow cover provides uniform surface conditions, whilst a constant distance between the transmitter and receiver antennas is achieved by mounting both on a single sledge. Geophysicists should take another critical care in acquiring georadar data, by keeping the system at accurate positioning. High-resolution investigations with measurement spacings of a few tens of centimetres require the use of automatic positioning systems (Lehmann and Green, 1999; Streich et al., 2006). In small areas (e.g. $50 \times 50 \text{ m}^2$) of particular interest, 3-D georadar experiments can be performed. This technique has proven to be highly successful in various applications (Grasmueck et al., 2004; Gross et al., 2004; Heincke et al., 2005) and is expected to work well on mountain permafrost. The foremost and the critical step is to sample the area of investigations so densely that one can obtain factual and unbiased subsurface images.

The study of Maurer and Hauck (2017) at 50 MHz antennas and a station spacing of 0.5 m to record georadar data along a $\sim 350 \text{ m}$ long profile provided very significant insights into the subsurface. Data were processed through a simple processing scheme by deploying automatic gain control, frequency filtering, spectral whitening, and F-X deconvolution. Topographic migration could not be done because of the unavailability of the detailed speed information along the profile. Instead, a simple time-to-depth conversion using a speed of $120 \text{ m}/\mu\text{s}$ (based on a common midpoint measurement in the central part of the rock glacier) is performed. It is reported that depth error associated with the overly simplified assumption of a homogeneous speed is difficult, but Maurer and Hauck (2017) adjudged that depth errors should not exceed 10%. The final section is displayed in the figure in Maurer and Hauck (2017). It is also observed that maximum depth penetration of the georadar signals is about 50 m. Several prominent dipping reflections extending over substantial portions of the profile are also reported by Maurer and Hauck (2017). They showed that a series of reflection signatures exists for the up-glacier part of the georadar section, which are most likely associated with debris and alternating snow accumulations due to the presence of the slope in formation that may have caused mass movement events from the top of the rock glacier. It is conspicuously shown that such features are quite common in georadar images of the upper part of rock glaciers; they represent the snow and debris sources that define the typical morphology of rock glacier bodies

(Berthling et al., 2003). Interestingly, the two dominant reflections originate from the ice-rich and moderately ice-rich layers (also observed in the borehole). The deeper reflection at $\sim 20\text{--}30$ m depth between 0 and 125 m distance corresponds to the bottom of the ice containing units and to the shear horizon observed in the inclinometer measurements described by Vonder Muhll and Holub (1992). Moreover, these measurements indicate displacement rates of 0.04 ma^{-1} at this depth.

8.12 GEOPHYSICAL IMAGING OF ALPINE ROCK GLACIERS BY CROSS-HOLE RADAR SURVEY: MURAGL CASE STUDY

As mentioned earlier, the Muragl rock glacier is also located in the Upper Engadine area. It is 100–300 m wide, is ~ 700 m long, and extends over an altitude range of 2,600–2,800 m (Maurer and Hauck, 2017). It is distinguished by rapid horizontal movements of up to 0.5 ma^{-1} (Kaab and Vollmer, 2000). It is observed that the rock glacier is substantially faster than typical rock glacier movements, which are of the order of several cma^{-1} . Like the Murtèl rock glacier, it exhibits pronounced flow lobes with transverse furrows and ridges. The irregular-flow lobe pattern of the Muragl rock glacier (Maurer and Hauck, 2017) may be the result of periodic mass movements, perhaps reflecting several generations of rock glacier evolution. Tomographic inversions of the travel time data were performed using an algorithm described by Maurer and Green (1997). Further details of the experiment can be found in Musil et al. (2006). Perspective images of the tomographic planes are explained by Maurer and Hauck (2017).

These are characterized by a background speed of $\sim 120\text{ mps}^{-1}$ and embedded zones of both higher and lower speed. At depths of <35 m, there are three zones of distinctly higher speed with values reaching 160 mps^{-1} (Maurer and Hauck, 2017), whereas in the depth interval 35–45 m, the speeds are as low as 100 mps^{-1} (Maurer and Hauck, 2017). They showed that the high-speed zone matched with the high-resistivity region, in which the sediments and boulder-based pores were filled with ice (Maurer and Hauck, 2017), whereas the underlying high-speed zone is located in a lower-resistivity region, in which the pores are mostly empty (Maurer and Hauck, 2017). These observations are entirely consistent with the relatively high radar speeds of ice (170 mps^{-1}) and air (300 mps^{-1}) and with the joint interpretations of georadar travel times and amplitudes (Musil et al., 2006). Since it is not possible to maintain such voids in an active rock glacier over long periods, they were probably filled with ice until quite recently (Musil et al., 2006).

Comparisons of the radar tomographic images with the results of inclinometer measurements performed in the borehole (Arenson et al., 2002) reveal an interesting similarity with the Murtèl rock glacier. As indicated by the black arrow in the figure in Maurer and Hauck (2017), most of the displacements are concentrated at the bottom of the ice-rich layer, defined by the high-speed zone. It is also reported that the much higher displacement rates at the Muragl rock glacier relative to those at the Murtèl rock glacier may be the result of generally higher temperatures close to the melting point of ice, as measured in the boreholes (Vonder Muhll et al., 2005).

8.13 ARCHAEOLOGICAL GEOPHYSICS

Archaeological geophysics is a range of techniques for the minimally invasive, remote investigation of the physical parameters of the near-surface environment. It is very important to undertake archaeological survey to get information about the stratigraphy of the survey area, locate anthropogenic traces of the past, document their spatial dimensions, and under ideal conditions, explore the physical properties of subsurface materials. It is known that both material culture items, such as a building's foundations, and indirect indications of anthropogenic activity, such as subsurface disturbance or evidence of burning, are excellent direct targets for geophysical investigations since they can be differentiated on the basis of their material properties from the wider soil context. With the GPR also deployed for the direct location of archaeological material, geophysical techniques can make an important contribution to geoarchaeological investigations by elucidating the

site stratigraphy and mapping its lateral geometry. It is observed that, in some cases, when locating prehistoric material buried offshore or within open Palaeolithic sites, the reconstruction of past landscapes may make a more important contribution to archaeological investigations than the direct geophysical detection of archaeological materials.

Excellent detailed introductions to the application of specific geophysical techniques to archaeology in general are available for ground-penetrating radar (GPR) (Manataki et al., 2015; Conyers, 2013, 2015), magnetometry (Armstrong and Kalayci, 2015; Aspinall et al., 2008), resistivity (Schmidt, 2013), and electromagnetic techniques (Simon and Moffat, 2015). The **GPR** was deployed for various purposes in India that include shallow subsurface sedimentary architecture of the Modwa spit at Gulf of Kachchh (Shukla et al., 2008); shallow cover over a coal mine at Nagpur (Soni et al., 2007); delineation of old mine workings in the Raniganj coalfield, West Bengal, India (Seguin et al., 1994); active faults at the western Gangetic plains (Bhosle et al., 2007); assessment of pollution near a landfill site in Nagpur (Pujari et al., 2007); fluvial geomorphology and neotectonic activity at Kachchh (Patidar et al., 2007); hydrogeological conditions related with mining activity (Singh, 2003); groundwater exploration in granitic hard rocks at Hyderabad (Sonkamble et al., 2013); and concealed thrusts in the middle **Gangetic** Plain (Pati et al., 2011).

Case Study: Australia and Greece

GPR applications in geoarchaeology range from the determination of the depth to bedrock to the mapping of palaeolandscape features (such as channels and shorelines), reported by Sarris et al. (2017), which located anthropogenic earthwork features like canals or moats, and the identification of soil disturbance due to habitation. Additionally, GPR allows mapping unmarked graves, which often can be identified through soil anomalies caused by the excavation of tombs. Unfortunately, the high attenuation of radar energy in salt water precludes the application of GPR in marine surveys, but it is suitable for use in limnic environments (lake overturn is a very rare type of natural disaster) (Sarris et al., 2017).

Case Study: Ladang Sungai Batu, Merbok, Kedah

Shokri and Amin (2016) performed an archaeological prospection at Lot 1736, Ladang Sungai Batu, Merbok, Kedah, as shown in the figure in Shokri and Amin (2016). The National Heritage Department (NHD), with the cooperation of several agencies, has discovered some stone structures that resemble a wall of a historical structure, as show in the figure in Shokri and Amin (2016). According to the NHD and local residents, there are 97 areas of potential historical sites identified, and nearly 50 sites were found. The scope of research in this case study is to conduct geophysical studies to help NHD carry out work on underground mapping and other methods that are appropriate to the field of archaeology. Accordingly, GPR detection was performed to determine the position and depth of real artefacts on 27 January 2016.

After the completion of data acquisition, the data was processed to obtain information such as identifying whether a historical structure exists or not in the area. The software used is 3-D ScanReflex for data processing. The function of this software is to combine all the data, such as data detection transversal-longitudinal, and thus, forming a 3-D shape of the area, the size depends on the investigated area. Several processing steps to filter data are DC shift, static correction (automatic), dewow, profit function, background removal, and migration. After performing data filtering, the 3-D model needs to be drafted. If there is a difference in amplitude in the images of each layer of soil, it will be known by the colour bar selected. Depending on the amplitude, the reflection on the objects contained in the basement can be determined. The figure in Shokri and Amin (2016) shows a sectional dimension of the study area. Referring to

the figure in Shokri and Amin (2016), there is a large difference in amplitude around a depth of 0.5 m to 1 m and 4 m to 5 m; the amplitude difference indicates that there is something under the soil layer. The amplitude changes for each hose of 0.5 m indicate that the soil conditions change for each layer. Shokri and Amin (2016) showed the picture of the real site after the excavation done by the NHD. They found impressions of the river wall, which was also corroborated by the findings of Iklil Izzati Zakaria and Abdullah (2011), and they stated the existence of an upriver–downriver economical interaction. It is reported that the discovery of these two river jetties affirmed the complexity of the site of Sungai Batu as a centre not only for religious and production activities but also the exchange of commodities, namely, the iron-based trade.

Case Study: GPR Application for a Historic Cemetery of Alabama (South-Eastern Region of the United States)

Ground-probing or ground-penetrating radar (GPR) is the most effective technique of innovation in ‘archaeological prospection’, which was first used in the USA in the early 1970s. Since GPR was originally designed to serve as an aid to engineering geology studies (Dolphin et al., 1978; Ulriksen, 1982; Batayneh et al., 2002), now it has become an important geophysical technique for mapping tool for archaeologists (Bevan, 1983; Vaughan, 1986; Imai et al., 1987). GPR is of limited use where either the soil or rocks are of high conductivities (Batayneh, 2011; Jones, 2008), showing a typical record obtained by means of a GPR system.

Application of GPR Survey in Ahichchhatra, Uttar Pradesh, India: Case Study

Sravanthi et al. (2012) carried out GPR prospecting in order to assess the extent of buried features across at least four sites of this vast city, two of them with high resolution (Prabhakar and Korisettar, 2017; courtesy: Bhuvan Vikrama). Ahichchhatra is dated from the 2nd millennium BCE to 14th century CE and has witnessed several phases of development and settlement expansion. The GPR method seems to be the only cost-effective, most suitable, and less time-consuming methodology for such a site when compared with excavation. The important aspect of GPR survey is the testing of anomalies obtained from GPR profiling through excavation.

Ground-Probing Radar (GPR) Investigation at the Harappan Site of Dholavira, in Kachchh, Gujarat: Case Study

Geophysical investigation at the Harappan site of Dholavira, in Kachchh, Gujarat, is a noteworthy example. Excavations at the site had revealed monumental architecture and water management systems. The GPR prospecting carried out during three site visits by scientists of IIT–Gandhinagar helped in understanding the vast buried remains in the unexcavated portions of the site (Majumdar, 2015; Agrawal, 2015). A vast and flat area to the east of East Reservoir and to the north–north–west of the Manhar River was also studied using GPR survey (Prabhakar and Korisettar, 2017). The study area at the historical Harappan site of Dholavira in Kachchh, Gujarat, India, that lies immediately to the south of the Lower Town was also considered for probing as it was covered with a sheet of water during flash floods and heavy rains; besides, the area is also flat and devoid of any prominent aboveground architectural features (Prabhakar and Korisettar, 2017), which was investigated by GPR and unravelled conspicuously the T-shaped parallel walls with a spacing of 7–8 m between them in an area with a north–south alignment, along with a spread of rubble, which may represent collapsed walls (Majumdar, 2015; Agrawal, 2015). The overall findings from the investigations indicated the presence of shallow

and smaller reservoirs to the east of East Reservoir, possibility check dams, and fallen walls forming rubble and scattered debris.

Integrated Ground-Penetrating Radar (GPR) and DGPS Method to Characterize the Litang Fault, China: Case Study

Most intriguingly, an integrated ground-penetrating radar (GPR) and Global Positioning System (GPS) survey has been used comprehensively to investigate the subsurface materials or buried objects with geographical information. Zhang et al. (2022) studied the Litang fault in China, where he evaluated the feasibility and efficiency of the integrated GPR and DGPS method. The 250 MHz and 500 MHz GPR profiles were covered in the four survey sites along the Litang fault.

The 250 MHz and 500 MHz GPR profiles in the four survey sites were processed using the RefexW7.2 software, as shown in the figure in Zhang et al. (2022). The subsurface structure on the GPR sections (Zhang et al., 2022) was not well displayed and identified without topographic correction. The figure in Zhang et al. (2022) shows the 250 MHz and 500 MHz images that were employed for topographic correction with the high-resolution topographic data. Compared to the GPR profiles in the figure in Zhang et al. (2022), the surface elevation of the survey line is well shown on the GPR images in the figure in Zhang et al. (2022), and the true subsurface structure is displayed on the GPR image as topographic variation. The subsurface structures can be interpreted in a better fashion for clear understanding of the target body, which can be achieved only using the GPR data after topographic correction.

A series of 250 MHz and 500 MHz GPR profiles were conducted, and the range of the survey lines was from 6 to 260 m. The comparison results between the number of the traces on the GPR images and the pulse events of the GPS signal receiver were shown in the figure in Zhang et al. (2022). The horizontal axis represents the number of the pulse events, and the vertical axis displays the number of GPR survey lines. In order to clearly show the difference in the values between the number of traces on the GPR images and the pulse events of the GPS signal receiver, different distances of the GPR survey lines at the bar charts were drawn in the figure in Zhang et al. (2022). The difference value of the pulse events, whether 250 MHz and 500 MHz GPR antennas, indicates that the differences slightly increase when the distance increases on the graphs. Furthermore, the study results demonstrate that the integrated GPR and DGPS method is suitable for collecting GPR data with the geographical coordinates for a continuous and long-distance survey in the rugged terrain.

Application of GPR in Locating Fresh Water in Coastal Terrain at Vizag Steel Plant (VSP), Visakhapatnam, Andhra Pradesh, India

It is observed that groundwater exploration can be carried out using the GPR technique, for which the geological formations of the area are classified into two types, namely, hard crystalline rocks and unconsolidated sediments. Khondalites, charnockites, feldspathic gneisses, and granites are grouped into hard crystalline rocks, whilst unconsolidated formations include red sediments, colluvium, alluvium, and coastal sands. The area is characterized by red loamy soil in the northern part and red sandy soils in the southern part. Coastal alluvial and sandy soils are seen in the coastal plain of the Vizag Steel Plant, Andhra Pradesh, India, where a patch of lateritic soil occurs in the north-eastern part along with tidal flats (mud flat), unconsolidated to semiconsolidated sand, silt, and clay materials, which are very much responsive to electromagnetic signals. In order to assess the groundwater conditions

at VSP and township, a GPR profile of 20 m length is carried out using 40 MHz antenna in time mode (continuous mode) (Dhakate et al., 2015). Radargram (GPR images) shows highly weathered formation up to a depth of 13 m, followed by saturated weathered formation up to a depth of 33 m. The study highlighted that more reflective signals were observed up to a depth of 13 m, indicating a highly weathered formation, followed by moderate reflection from 13 m to 33 m, suggestive of the weathered formation (Dhakate et al., 2015). Continuous discharge measurements were recorded after the drilling of borewell, and it was found that the borewell yielded about 104.7 gal/h of water. It is found that the radargram shows highly weathered formations up to a depth of 20 m, because of more reflected signals at this depth, followed by weathered/fractured formation up to a depth of 38 m due to moderate reflected signals (Dhakate et al., 2015). The combined geophysical data interpretation is advantageous over the application of a single-method data interpretation for pinpointing fresh groundwater resources in a saline-affected area.

Application of GPR for Groundwater in Granitic Terrain, Hyderabad, India: Case Study

The GPR has become an important tool in the investigation of groundwater resources (Liu and Li, 2001; Carreon-Freyre and Cerca, 2006). In this study, the usefulness of GPR in characterizing a groundwater-enriched lineament zone in the granite exposure near Gajularamaram in the north-western part of Hyderabad City (Maheswari et al., 2013) is described. The lineament-parallel fractures are quite abundant in the WNW part of the lineament but are less pronounced in the ESE part, where subhorizontal to gently dipping sheet joints are observed on either sides of the lineament (Maheswari et al., 2013). Some of the sheet joints show the discharge of groundwater (Maheswari et al., 2013).

The final processed GPR sections are shown in the figures in Maheswari et al. (2013), respectively. The processed GPR sections provide the subsurface geological information up to a depth of ~35 m for 100 MHz and ~6 m for 400 MHz frequencies. The radargram reveals the thickness and internal structure of the soils, weathered zones, fresh granite bedrock, fracture/fault patterns, and groundwater signatures. Radar reflectors characterize these geological features and materials. The GPR sections (profiles 1 and 2) show two layers, a 15 m thick upper layer and a >15 m thick lower layer (Maheswari et al., 2013). The lower layer is interpreted as the fresh granite bedrock, while the upper layer is weathered granite with a thin soil cover. An undulating reflector characterizes the contact between the soil cover and the bedrock at the depth range of 5 m to 15 m. Near the north-eastern end of these profiles, the GPR sections reveal a 10 m wide vertical anomaly zone (where the prominent reflectors are absent), which is interpreted as vertical fracture/fault zone with or without intense weathering. The presence of a strong reflector near the upper surface of the lineament zone may have originated from the upper surface of the groundwater table in the lineament. These areas are also characterized by the springs. Geological problems with many near-vertical parallel lineaments as well as sheet joints can also be resolved by deploying GPR methods to detect and delineate those vertical lineaments located at different distances in the different parts of the study area (Maheswari et al., 2013).

The GPR sections (Maheswari et al., 2013) revealed subhorizontal to near-vertical features. Many prominent vertical lineaments are observed around 10 m, 40 m, 65 m, 75 m, 85 m, and 180 m, respectively. In addition, a few minor lineaments are also present. Near the surface, soil cover of variable thickness (0.5 to 5 m) is observed. Sometimes, many subhorizontal reflectors at a depth of 10–20 m can be interpreted as sheet joints similar to those in the nearby quarry exposures (Maheswari et al., 2013). The 100 MHz GPR section (Maheswari et al., 2013) revealed several gently dipping sheet joints throughout the section. In addition, the 400 MHz

GPR data (Maheswari et al., 2013) clearly showed thickening of the soil cover from the SSW end of the profile from the lineament boundary to the central part of the lineament. Soil thickness is ~ 25 cm near the SSW end, and it increases to ~ 1 m at the NNE end of the profile, as reported by Maheswari et al. (2013).

The GPR study establishes the presence of the lineament in the near surface. The geological and GPR observations point out the presence of subvertical fractures and faults in the granite bedrock. The shallow radar reflections over the lineament in profiles 1 and 2 show the groundwater table at or near the surface. The lineament in the study area is therefore a classic example of a typical groundwater reservoir in hard rock terrains. The GPR study can also be integrated with other techniques, such as multichannel analysis of surface waves (MASW), which provides shear wave velocity image of the shallow subsurface (Seshunarayana et al., 2008a, 2008b), and soil–gas emanometry (Reddy et al., 2010).

8.14 GEOTECHNICAL STUDIES

Under electromagnetic surveys, geotechnical site characterization has become a very important deliverable, resulting in using ground-penetrating radar (GPR). During the course of a typical GPR survey, these tools are used to measure spatial variations in the travel times and magnitudes of pulsed EM radiation that has been reflected from subsurface features (generally geologic boundaries) of interest. The 2-D GPR profile presented by Anderson et al. (2007) consisted of multiple adjacent traces (reflection amplitude plotted as a function of two-way travel time) which were acquired at predetermined intervals along a 2-D traverse across a shallow stream. The beauty of the GPR profile is that it contains only one prominent reflection, representing a simple two-layered (water overlying relatively uniform sand) model.

The EM velocities are dictated by the dielectric constants assigned to each of the two layers identified on the GPR profile (Anderson et al., 2007). The arrival time of the reflected event (water–sand interface) at any trace location on the GPR profile is a function of the EM velocity of water, and the amplitude of the GPR reflection at any trace location is a function of the contrasting dielectric constants of water and sand. Properly acquired and processed geotechnical-geophysical survey data can generally be transformed into a physical property model. GPR data are frequently transformable into corresponding 2-D or 3-D ‘reflection amplitude constant/depth’ model, as shown in Table 2.1 of Chapter 2 (Anderson et al., 2007), depicting one reflecting horizon (water–sand interface). It is seen that the ‘dielectric constant–depth’ model and the vertical ‘time–depth’ scale can be transformed into a vertical ‘depth’ scale (time-to-depth conversion). Alternative physical property models could be in the form of ‘EM velocity–depth’, ‘EM velocity–time–depth’, etc. If additional geophysical or geological (including hydrologic, engineering, mining) constraints are available, then physical property models can be transformed into ‘typical site models’ (Table 2.1, Chapter 2). A typical site model for a geotechnical geophysical investigation is a geologic model describing many soil or rock properties of the study area. The GPR physical property model has been transformed into a simple hydrologic-geologic model on the basis of site geomorphology and subsurface geologic control, as described by Anderson et al. (2007).

APPENDIX 1

These formulae can be deduced based on the standard principle of electromagnetics, for which if we consider a field generated by a horizontally directed current dipole, located in a homogeneous unbounded conductive medium with a conductivity σ and a magnetic permeability equal to that of free space μ_0 , energized with a current impulse that has the form of a delta function, it is very well-known that the vertical component of the magnetic field of a harmonic dipole source with the unit horizontal moment $p_x = 1$ is

$$H_z(\omega) = -\frac{1}{4\pi} \frac{\partial}{\partial y} \left(\frac{e^{ikr}}{r} \right) \quad (\text{a1.8})$$

where r is the distance from the transmitter to the receiver, ω is frequency, and k is a wave number. Applying to this expression a Fourier transform from the frequency domain to the time domain, we find for a magnetic field excited by a delta current pulse the following expression:

$$H_z^\delta(t) = -\frac{1}{4\pi} \frac{\partial}{\partial y} \left[\frac{1}{r} \left(\frac{1}{2\pi} \int_{-\infty}^{\infty} e^{ikr} e^{-i\omega t} d\omega \right) \right] \quad (\text{a2.8})$$

where t is time. After some algebra, this equation can be rewritten as (Zhdanov and Keller, 1994)

$$H_z^\delta(t) = -\frac{1}{4\pi} \frac{\partial}{\partial y} \left[\frac{1}{r} \frac{8\pi^2 r (2\pi)^{\frac{1}{2}}}{\mu_0 \tau^3 \sigma} e^{-2\left(\frac{\pi r}{\tau}\right)^2} \right] = \frac{8\pi^3 (2\pi)^{\frac{1}{2}}}{\mu_0 \sigma} \frac{y}{r^5} e^{-2\left(\frac{\pi r}{\tau}\right)^2} \quad (\text{a3.8})$$

where the time domain parameter τ is defined as a combination of various constants that group together naturally:

$$\tau = \frac{2\pi}{\left(\frac{\mu_0 \sigma}{2t} \right)^{\frac{1}{2}}} \quad (\text{a4.8})$$

It is evident that this parameter has the dimension of length and is usually calculated to be

$$\tau = \left(2\pi \rho t \times 10^7 \right)^{\frac{1}{2}} \quad (\text{a5.8})$$

We see from the expressions a3.8 that the characteristic decline of the field from an impulsive source as a function of distance from the source is determined by the exponential multiplier $e^{-2\left(\frac{\pi r}{\tau}\right)^2}$, in which r appears as a ratio to the length of τ . This ratio length to τ is a harmonically driven source, which is called the electrical distance:

$$r_\tau = \frac{r}{\tau} \quad (\text{a6.8})$$

Note that τ plays the same role as does the wavelength λ in the frequency domain. As is the case in the frequency domain, we can recognize three zones of behaviour for the fields: (1) the near zone, for which $r \ll \tau$; (2) the intermediate zone, in which $r \approx \tau$; and (3) the far zone, for which $r \gg \tau$.

There is a trade-off between time and distance in defining the various zones, which means the far zone can be associated with the early part of the transient response and the near zone can be associated with the latter part of the transient waveform. Since the reciprocity involves in the roles played by time and distance, the far zone is therefore sometimes called the 'early time zone', and the near zone is called the late time zone. These observations suggest that the near field corresponds to late times because the late times have a lower-frequency content that extends the near field farther from

the transmitter. Consequently, at very early times, almost everything is in the far field because this corresponds to high frequencies and very short electrical distances.

Let us turn now to the characteristic features of the transient fields in the near zone, $r \ll \tau$. In equation a3.8, taking a limit $\frac{r}{\tau} \rightarrow 0$, we have

$$H_z^\delta(t) \approx \frac{8\pi^3 (2\pi)^{\frac{1}{2}}}{\mu_0} \frac{\rho y}{r^5} \quad (\text{a7.8})$$

where $\rho = \frac{1}{\sigma}$ is resistivity. The same expression for the magnetic field in the near zone caused by a harmonic source can be written as:

$$H_z(\omega) \approx \frac{1}{4\pi} \frac{y}{r^3} e^{-i\omega t} \quad (\text{a8.8})$$

where $r = \sqrt{x^2 + y^2 + z^2}$ and where x, y, z are the coordinates of the receiver in the Cartesian system, with the origin in the location of the electric dipole source. Comparing formulas a7.8 and a8.8, we see that although $H_z(\omega)$ is independent of the resistivity of the medium, the time domain signal $H_z^\delta(t)$ is linearly proportional to ρ . Moreover, if we analyze $H_z^\delta(t)$ and $|H_z(\omega)|$ as functions of depth z for fixed values of x_0, y_0, t_0 , and ω_0 , we can see that in the near zone, the field of a harmonic dipole has reciprocal relation with depth, and it found decreases with an increase of depth, whilst the transient field is found to remain uniform with depth. This provides the possibility, in theory at least, of investigating the geoelectric section to great depths with observations made close to the source. In practice, it is only required that a sufficiently strong pulse be applied to the source to explore to very great depths.

REFERENCES

- Agrawal, S., 2015. GPR data analysis of weak signals using modified S-transform. Unpublished M Tech thesis. Indian Institute of Technology, Gandhinagar.
- Aitken, M., 1974. *Physics and Archaeology*. 2nd edn. Oxford: Clarendon Press.
- Al-Zoubi, A., Haddadin, G. and Batayneh, A., 1998. Geophysical methods for locating a prehistoric site in the Al Hummaymeh area, southwest Jordan. *Journal of Environmental and Engineering Geophysics*, vol. 3, pp. 157–161.
- Anderson, N.L., Ismail, A.M. and Thitimakorn, T., 2007. Ground-penetrating radar: A tool for monitoring bridge scour. *Environmental and Engineering Geoscience*, vol. 13, p. 10.
- Annan, A.P. and Chua, L.T., 1992. Ground penetrating radar performance predictions. *The Geological Survey of Canada*, vol. 90, no. 4, pp. 5–13.
- Annan, A.P., Waller, W.M., Strangway, D.W., Rossiter, J.R., Redman, J.D. and Watts, R.D., 1975. The electromagnetic response of a low loss, 2-layer, dielectric earth for horizontal electric dipole excitation. *Geophysics*, vol. 40, pp. 285–296.
- Annan, A.P. 2009. Electromagnetic principles of ground penetrating radar. In: Jol HM (ed) *Ground penetrating radar: theory and applications*, 2009. Elsevier, Solvenia, p. 440.
- Archie, G.F., 1942. The electrical resistivity log as an aid in determining some reservoir characteristics. *Transactions AIME*, vol. 146, pp. 54–62.
- Arcone, S.A., Lawson, D.E., Delaney, A.J., Strasser, J.C. and Strasser, J.D., 1998. Ground penetrating radar reflection profiling of groundwater and bedrock in an area of discontinuous permafrost. *Geophysics*, vol. 63, no. 5, pp. 1573–1584.
- Arenson, L., Hoelzle, M. and Springman, S., 2002. Borehole deformation measurements and internal structure of some rock glaciers in Switzerland. *Permafrost and Periglacial Processes*, vol. 13, no. 2, pp. 117–135.

- Armstrong, K.L. and Kalayci, T., 2015. Images of the past: Magnetic prospection in archaeology. In: Sarris, A. (ed), *Best Practices of Geoinformatic Technologies for the Mapping of Archaeolandscape*. Oxford: Archaeopress Publishing, pp. 1–12.
- Aspinall, A., Gaffney, C. and Schmidt, A., 2008. *Magnetometry for Archaeologists*. Lanham: AltaMira Press.
- Auken, E., Christiansen, A.V., Jacobsen, B.H., Foged, N. and Sorensen, K.I., 2005. Piece-wise 1D laterally constrained inversion of resistivity data. *Geophysical Prospecting*, vol. 53, pp. 497–506, DOI: 10.1111/j.1365-2478.2005.00486.x.
- Auken, E., Christiansen, A.V., Jacobsen, L.H. and Sorensen, K.I., 2008. A resolution study of buried valleys using laterally constrained inversion of TEM data. *Journal of Applied Geophysics*, vol. 65, no. 1, pp. 10–20.
- Banos, A., 1966. *Dipole Radiation in the Presence of a Conducting Half-Space*. New York: Pergamon Press.
- Batayneh, A.T., 2011. Archaeogeophysics—archaeological prospection—a mini review. *Journal of King Saud University a—A Science*, vol. 23, no. 1, pp. 83–89.
- Batayneh, A.T., in press. The use of magnetometry and pole—dipole resistivity for locating Nabataean Hawar archaeological site in the SW-Jordan. *Archaeological and Anthropological Sciences Journal*.
- Batayneh, A.T., Abueladas, A. and Moumani, K., 2002. Use of ground penetrating radar for assessment of potential sinkhole conditions: An example from Ghor al Haditha area. *Jordan Environmental Geology*, vol. 41, pp. 977–983.
- Batayneh, A.T., Khataibeh, J., AlRshdan, H., Tobasi, U. and Al-Jahed, N., 2006. The use of microgravity, magnetometry and resistivity surveys for the characterization and preservation of an archaeological site of Umm er-Rasas. *Jordan Archaeological Prospection*, vol. 14, pp. 60–70.
- Bauer-Gottwein, P., Gondwe, B.N., Christiansen, L., Herckenrath, D., Kgotlhang, L. and Zimmermann, S., 2010. Hydrogeophysical exploration of three-dimensional salinity anomalies with the time-domain electromagnetic method (TDEM). *Journal of Hydrology*, vol. 380, no. 3–4, pp. 318–329.
- Beamish, D., 1998. Three-dimensional modelling of VLF data. *Journal of Applied Geophysics*, vol. 39, no. 2, pp. 63–76.
- Benson, R.C., Glaccum, R.A. and Noel, M.R., 1983. *Geophysical Techniques for Sensing Buried Wastes and Waste Migration*. Westerville, OH: National Water Well Association.
- Berthling, I., Etzelmüller, B., Wale, M. and Sollid, J.L., 2003. Use of Ground Penetration Radar (GPR) soundings for investigating internal structures in rock glaciers: Examples from Prins Karls Forland, Svalbard. *Z. Geomorphol. Supplementband*, vol. 132, pp. 103–121.
- Bevan, B., 1983. Electromagnetics for mapping buried earth features. *Journal of Field Archaeology*, vol. 10, pp. 47–54.
- Bhosle, B., Parkash, B., Awasthi, A.K., Singh, V.N. and Singh, S., 2007. Remote sensing-GIS and GPR studies of two active faults, Western Gangetic Plains, India. *Journal of Applied Geophysics* (Elsevier, B.V.), Amsterdam, Netherlands, ScienceDirect, vol. 61, pp. 155–164.
- Bosch, F.P. and Gurk, M., 2000. Comparison of RF-EM, RMT and SP measurements on a karstic terrain in the Jura Mountains (Switzerland). The Proceedings of the Seminar Electromagnetische Tiefenforschung, organized by the Deutsche Geophysikalische Gesellschaft (DGG), Berlin, Germany, pp. 51–59.
- Brekovskikh, L.M., 1980. *Waves in Layered Media*. 2nd edn. New York: Academic Press.
- Brewster, M.L. and Annan, A.P., 1994. Ground-penetrating radar monitoring of a controlled DNAPL release: 200 MHz radar. *Geophysics*, vol. 59, pp. 1211–1221.
- Brewster, M.L., Annan, A.P., Greenhouse, J.P., Kueper, B.H., Olhoeft, G.R., Redman, J.D. and Sander, K.A., 1995. Observed migration of a controlled DNAPL release by geophysical methods. *Ground Water*, vol. 33, pp. 977–987.
- Burger, H.R., Sheehan, A.F. and Jones, C.H., 2023. *Introduction to Applied Geophysics*. New York: Publisher W.W. Norton, Cambridge University Press, p. 600.
- Cagniard, L., 1953. Basic theory of the magnetotelluric method of geophysical prospecting. *Geophysics*, vol. 18, pp. 605–645.
- Candansayar, M.E. and Tezkan, B., 2006. A comparison of different radiomagnetotelluric data inversion methods for buried waste sites. *Journal of Applied Geophysics*, vol. 58, pp. 218–231.
- Carreon-Freyre, D.C. and Cerca, M., 2006. Delineating the near surface geometry of the fracture system affecting the Queretaro valley, Mexico: Correlation of GPR signatures and physical properties of sediments. *Near Surface Geophysics*, pp. 49–55.

- Chen, J. and Raiche, A., 1998. Inverting AEM data using a damped Eigen parameter method. *Exploration Geophysics*, vol. 29, no. 2, pp. 128–132, DOI: 10.1071/EG998128.
- Chen, W., Xue, G., Khan, M.Y. and Li, H., 2017. Using SOTEM method to detect BIF bodies buried under very thick and conductive quaternary sediments, Huoqiu Deposit, China. *Pure and Applied Geophysics*, vol. 174, no. 3, pp. 1013–1023.
- Conyers, L.B., 2013. *Ground-Penetrating Radar for Archaeology*. Lanham: AltaMira Press.
- Conyers, L.B., 2015. *Ground-Penetrating Radar for Geoarchaeology*. London: Wiley.
- Dalan, R.A., 2007. A review of the role of magnetic susceptibility in archaeogeophysical studies in the USA: Recent developments and prospects. *Archaeological Prospection*, vol. 15, no. 1, pp. 1–31.
- Dalan, R.A. and Banerjee, S.K., 1998. Solving archaeological problems using techniques of soil magnetism. *Geoarchaeology*, vol. 13, no. 1, pp. 3–36.
- Daniels, J.J., 1996. *Surface Penetrating Radar*. London: The Institution of Electrical Engineers.
- Davis, J.L. and Annan, A.P., 1989. Ground-penetrating radar for high-resolution mapping of soil and rock stratigraphy. *Geophysical Prospecting*, vol. 37, pp. 531–551.
- Dhakate, R., Kumar, V.S., Amarender, B., Sankaran, S. and Rao, V.V.S.G., 2015. Integrated geophysical and hydrochemical approach for locating fresh water locations in a coastal terrain. *Water Resources Management*, vol. 29, pp. 3401–3417.
- Dinanta, P., Wicaksono, N., Hidayat, W., Ramadhan, R., Noor, M.R., Cassidy, D., Sudiyanto, Y., Herald, E., Afifuddin and Al Ghiffary, M.R., 2023. Case study of ground penetration radar (GPR) to assess lead migration. *Results in Geophysical Sciences* (Science Direct), vol. 14, p. 100055.
- Dolphin, L., Tanzi, J. and Beatty, W., 1978. Radar probing of Victorio Peak. *New Mexico Geophysics*, vol. 43, pp. 1441–1448.
- Dupis, A. and Choquier, A., 1996. How electrostatic was introduced in subsurface geophysics. *Proceedings, EEGS, Nantes*, pp. 125–128.
- Duque, C., Calvache, M.L., Pedrera, A., Martin-Rosales, W. and Lopez-Chicano, M., 2008. Combined time domain electromagnetic soundings and gravimetry to determine marine intrusion in a detrital coastal aquifer (Southern Spain). *Journal of Hydrology*, vol. 349, no. 3–4, pp. 536–547.
- Ekko_Project Team, 2015. *Ekko Project User's Guide (2015) Instrument Manual*. Mississauga, ON: Sensors & Software Inc. 2011-00125-11.
- El Assel, N., Kchikach, A., Teixidó, T., Peña, J.A., Jaffal, M., Guéring, R., Lutz, P., Jourani, E. and Amaghaz, M., 2011. A ground penetrating radar and electrical resistivity tomography prospection for detecting sterile bodies in the phosphatic bearing of Sidi Chennane (Morocco). *International Journal of Geosciences*, vol. 2, pp. 406–413.
- Ellis, R.G., 1998. Inversion of airborne electromagnetic data. *Exploration Geophysics*, vol. 29, pp. 121–127, DOI: 10.1071/EG998121.
- Farquharson, C.G., Oldenburg, D.W. and Routh, P.S., 2003. Simultaneous 1D inversion of loop-loop electromagnetic data for magnetic susceptibility and electrical conductivity. *Geophysics*, vol. 68, pp. 1857–1869, DOI: 10.1190/1.1635038.
- Foster, E., 1968. Further developments of the pulsed induction metal detector. *Prospezioni Archeologiche*, vol. 3, pp. 95–99.
- Fountain, D., 1998. Airborne electromagnetic systems—50 years of development. *Exploration Geophysics*, 29, 1–11, DOI: 10.1071/EG998001.
- Franklin, J.M., H.L. Gibson, H.L., Jonasson, I.R. and Galley, A.G., 2005. Volcanogenic massive sulphide deposits. *Econ. Geol.*, pp. 523–560, 100th Anniversary vol. 523–560.
- French, C., 2003. *Geoarchaeology in Action. Studies in Soil Micromorphology and Landscape Evolution*. London: Routledge.
- Frischknecht, F.C., 1967. Fields about an oscillating magnetic dipole over a two-layer earth, and application to ground and airborne electromagnetic surveys. *Quarterly of the Colorado School of Mines*, 62, no. 1, 1–326.
- Fullagar, P.K. and Reid, J.E., 2001. Emax conductivity-depth transformation of airborne TEM data. 15th Geophysical Conference and Exhibition, Australian Society of Exploration Geophysicists, Hobart, Tasmania.
- Giovanni, L., 2012. Ground penetrating radar: A useful tool for shallow subsurface stratigraphy characterization. In: Dr. Elitok, Ö. (ed), *Stratigraphic Analysis of Layered Deposits*. Italy: Rome, Institute of Archaeological and Monumental Heritage, National Council of Research – (CNR-IBAM).
- Gogineni, S., Chuah, T., Allen, C., Jezek, K. and Moore, R.K., 1998. An improved coherent radar depth sounder. *Journal of Glaciology*, vol. 44, no. 148, pp. 659–669.

- Grasmueck, M., Weger, R. and Horstmeyer, H., 2004. Three dimensional ground-penetrating radar imaging of sedimentary structures, fractures, and archaeological features at submeter resolution. *Geology*, vol. 32, no. 11, pp. 933–936.
- Gross, R., Green, A.G., Horstmeyer, H. and Begg, J.H., 2004. Location and geometry of the Wellington Fault (New Zealand) defined by detailed three-dimensional georadar data. *Journal of Geophysical Research*, vol. 109, no. B5, p. B05401, DOI: 10.1029/2003JB002615.
- GSI, 2015. Central Head Quarter (CHQ) e News. *Geological Survey of India (GSI)*, Published by Kolkata GSI Headquarters, vol. 46, no. 1, January–June.
- Guerin, R., Descloitres, M., Talbi, A., Coudrain, A. and Gallaire, R., 2001. Geophysical surveys for identifying saline groundwater in the semi-arid region of the central Altiplano, Bolivia. *Hydrologic Processes*, vol. 15, pp. 3287–3301.
- Guo, Z., Xue, G., Liu, J. and Wu, X., 2020. Electromagnetic methods for mineral exploration in China: A review. *Ore Geology Reviews*, vol. 118, p. 103357.
- Hauck, C., Guglielmin, M., Isaksen, K. and Vonder Muhll, D., 2001. Applicability of frequency domain and time-domain electromagnetic methods. *Permafrost and Periglacial Processes*, vol. 12, no. 1, pp. 39–52.
- Heincke, B., Green, A.G., van der Kruk, J. and Horstmeyer, H., 2005. Acquisition and processing strategies for 3D georadar surveying a region characterized by rugged topography. *Ceophysics*, vol. 70, no. 6, pp. K53–K61.
- Herckenrath, D., 2012. Informing groundwater models with near-surface geophysical data. Ph.D. Thesis. Department of Environmental Engineering, Technical University of Denmark.
- Hobbs, B., Li, G., Clarke, C. and Linfoot, J., 2005. Inversion of multi-transient electromagnetic data. 68th Conference & Technical Exhibition, EAGE, Extended Abstracts, A015, Vienna, Austria. Red Hook, NY: Curran Associates, Inc.
- Hoekstra, P. and Delaney, A., 1974. Dielectric properties of soils at UHF and microwave frequencies. *Journal of Geophysical Research*, vol. 79, pp. 1699–1708.
- Hollier-Larousse, R., Lagabriele, R. and Levillain, J.P., 1994. Utilisation de la radio magnetotellurique pour la reconnaissance en site aquatique. *Journal of Applied Geophysics*, vol. 31, pp. 72–84.
- Hubbard, S.S., Petersen, J.E., Majer, E., Zawislanski, T.P., Williams, K.H., Roberts, J. and Wobber, F., 1997. Estimation of permeable pathways and water content using tomographic radar data. *The Leading Edge*, vol. 16, pp. 1623–1628.
- Iklil Izzati Zakaria, M.S. and Jeffrey Abdullah, J., 2011. Ancient Jetty at Sungai Batu complex, Bujang Valley, Kedah. Presented at the Postgraduate Student Forum 2011, Current Asia Anthropology, Hong Kong.
- Imai, T., Sakayama, T. and Kanemori, T., 1987. Use of ground probing radar and resistivity surveys for archaeological investigations. *Geophysics*, vol. 52, pp. 137–150.
- Interpex Limited, 1990. *TEMIX User's Manual*. Golden, CO: Interpex Limited.
- Isaksen, K., Odegard, R.S., Eiken, T. and Soliid, J.L., 2000. Composition, flow and development of two tongue-shaped rock glaciers in the permafrost of Svalbard. *Permafrost and Periglacial Processes*, vol. 11, no. 3, pp. 241–257.
- Jayaram, B.N., 1982. Base metal mineralization in the Aladahalli schist belt, Karnataka', *Bull. Of G.S.I., series A, Ecological Geology*, vol. 48, p. 83, with annexures p. 41.
- Jones, G., 2008. Geophysical mapping of historic cemeteries. *Technical Briefs in Historical Archaeology*, vol. 3, pp. 25–38.
- Kaab, A. and Vollmer, M., 2000. Surface geometry, thickness changes and flow fields on permafrost streams: Automatic extraction by digital image analysis. *Permafrost and Periglacial Processes*, vol. 11, no. 4, pp. 315–326.
- Kaab, A. and Weber, M., 2004. Development of transverse ridges on rock glaciers: Field measurements and laboratory experiments. *Permafrost and Periglacial Processes*, vol. 15, pp. 15–379.
- Kaufman, A.A., 1989. A paradox in geoelectromagnetism, and its resolution, demonstrating the equivalence of frequency and transient domain methods. *Geoexploration*, vol. 25, no. 4, 287–317, DOI: 10.1016/0016-7142(89)90003-3.
- Kaufman, A.A. and Keller, G.V., 1983. *Frequency and Transient Soundings*. Amsterdam: Elsevier.
- Kaufman, A.A. and Morozova, G.M., 1970. Theoretical basis for the transient sounding method in the near-zone, Nauka.

- Kearey, P. and Brooks, M., 2002. *An Introduction to Geophysical Exploration*. Oxford: Blackwell Science.
- Keller, G.V., 1968. Electrical prospecting for oil. *Quarterly of the Colorado School of Mines*, vol. 63, no. 2, pp. 1–268.
- Kirsch, R., 2006. Umweltgeophysik in der Praxis: Untersuchung von Altablagerungen und kontaminierten Standorten, Script. Groundwater protection: Mapping of contaminations. In: Kirsch, R. (ed), *Groundwater Geophysics*. Berlin, Heidelberg: Springer.
- Kong, F.N., Bhasin, R. and Chadha, N., 2012. Development of handheld step frequency GPR using Fieldfox network analyser and its applications. Proceedings of Indian Geotechnical Conference, Delhi (Paper no. A161.).
- Lagabriele, R., 1986. Les bases de la radio-magnetotellurique. *Revue de l'industrie minerale, mines et carrieres*, October, pp. 373–384.
- Lamontagne, Y., Lodha, G., Macnae, J. and West, G.F., 1978. Towards a deep penetration EM system. *Bulletin of the Australian Society of Exploration Geophysics*, vol. 9, 12–21.
- Lampe, B., Holliger, K. and Green, A.G., 2003. A finite difference time-domain simulation tool for ground-penetrating radar antennas. *Geophysics*, vol. 68, pp. 971–987.
- Langevin, C.D., Fitterman, D. and Deszcz-Pan, M., 2003. Calibration of a variable-density groundwater flow model using detailed airborne geophysical data. Second International Conference on Saltwater Intrusion and Coastal Aquifers—Monitoring, Modeling and Management, Merida, Mexico.
- Lehmann, F. and Green, A.G., 1999. Semi-automated georadar acquisition in three dimensions. *Geophysics*, vol. 64, pp. 64–719.
- Lehmann, F. and Green, A.G., 2000. Topographic migration of georadar data: Implications for acquisition and processing. *Geophysics*, vol. 65, no. 3, pp. 836–848.
- Li, X., Bai, D.H., Yan, Y.L. et al., 2016. Three-dimensional inversion of magnetotelluric resistivity model with galvanic distortion. *Chinese Journal of Geophysics*, vol. 59, no. 6, pp. 2302–2315 (in Chinese).
- Liu, L. and Li, Y., 2001. Identification of liquefaction and deformation features using ground penetrating radar in the New Madrid seismic zone, USA. *Journal of Applied Geophysics*, vol. 47, pp. 199–215.
- Lu, H.Z., Liu, Y., Wang, C., Xu, Y. and Li, H., 2003. Mineralization and fluid inclusion study of the Shizhuyuan W-Sn-Bi-Mo-F skarn deposit, Hunan Province China. *Economic Geology*, vol. 98, no. 5, pp. 955–974.
- Macnae, J., King, A., Stolz, N., Osmakoff, A. and Blaha, A., 1998. Fast AEM data processing and inversion. *Exploration Geophysics*, vol. 29, pp. 163–169, DOI: 10.1071/EG99816.
- Madden, R. and Mackie, R., 1989. Three-dimensional magnetotelluric modeling and inversion. *Proceedings of the IEEE*, vol. 77, no. 2, pp. 318–333.
- Maheswari, K., Kumar, P.S., Mysaiah, D., Ratnamala, K., Sri Hari Rao, M. and Seshunarayana, T., 2013. Ground penetrating radar for groundwater exploration in granitic terrains: A case study from Hyderabad. *Journal Geological Society of India*, vol. 81, pp. 781–790.
- Majumdar, M., 2015. Bed rock profile analysis and interpretation of archaeological features using GPR at Dholavira. Unpublished M Tech thesis. Indian Institute of Technology, Gandhinagar.
- Manataki, M., Sarris, A., Donati, J.M., Garcia, C.C. and Kalayci, T., 2015. GPR: Theory and practice in archaeological prospection. In: Sarris, A. (ed), *Best Practices of Geoinformatic Technologies for the Mapping of Archaeolandscape*. Oxford: Archaeopress Publishing, pp. 13–24.
- Masini, N., Persico, R. and Rizzo, E., 2010. Some examples of GPR prospecting for monitoring of the monumental heritage. *Journal of Geophysics and Engineering*, vol. 7, no. 2, p. 190. DOI: 10.1088/1742-2132/7/2/S05.
- Maurer, H.R. and Hauck, C., 2017. *Geophysical Imaging of Alpine Rock Glaciers*. vol. 53, no. 180. Cambridge: Published online by Cambridge University Press.
- Maurer, H.R. and Green, A.G., 1997. Potential coordinate mislocations in cross-hole tomography: Results from the Grimsel Test Site, Switzerland. *Geophysics*, vol. 62, pp. 1696–1706.
- McNeill, J.D., 1990. Use of electromagnetic methods for groundwater studies. In: Ward, S.H. (ed), *Geotechnical and Environmental Geophysics*. vol. 1. Tulsa, OK: Society of Exploration Geophysicists, pp. 191–218.
- Meju, M.A., Fontes, S.L., Oliveira, M.F.B., Lima, J.P.R., Ulugergerli, E.U. and Carrasquilla, A., 1999. Regional aquifer mapping using combined VES-TEM-AMT/EMAP methods in the semi-arid eastern margin of Parnaiba-Basin, Brazil. *Geophysics*, inprint.
- Mishra, O.P. and Zhao, D., 2003. Tomography of northeast Japan forearc and its implications for interplate seismic coupling. *Geophysical Research Letters*, vol. 30, no. 16, p. 1850, DOI: 10.1029/2003GL017736.
- Moorman, B.J., Robinson, S.D. and Burgess, M.M., 2003. Imaging periglacial conditions with ground penetrating radar. *Permafrost and Periglacial Processes*, vol. 14, no. 4, pp. 319–329.

- Morrison, H.F. and Nichols, E.A., 1997. Mineral exploration with natural electromagnetic fields. 4th Decennial International Conference on Mineral Exploration, Proceedings, Houston, TX, pp. 527–54.
- Muller, I., 1983. Anisotropic properties of rocks detected with electromagnetic VLF (Very Low Frequency). Int. Symp. Field Measurements in Geomechanics, Zurich, September, Spec. Publ., pp. 273–282.
- Musil, M., Maurer, H.R., Holligerand, K. and Green, A.G., 2006. Internal structure of an alpine rock glacier based on crosshole georadar traveltimes and amplitudes. *Geophysical Prospecting*, vol. 54, pp. 54–273.
- Nabighian, M.N. and Macnae, J.C., 1987. Time domain electromagnetic prospecting methods. In: Nabighian, M.N. (ed), *Electromagnetic Methods in Applied Geophysics*. Tulsa, OK: Society of Exploration Geophysicists.
- Naskar, D.C., Das, L.K., Roy, K.K., Majumdar, R.K. and Choudhury, K., 2012. A re-examination of the crustal structure across Narmada-Son-Lineament (NSL), Madhya Pradesh, Central India using magnetotelluric and gravity modelling. *Journal of Geophysics*, vol. 33, no. 3, pp. 69–77.
- Obukhov, G.G., 1968. About some properties of the non-stationary electromagnetic fields in the earth and their applications in electrical prospecting. *Izvestia, Physics of the Earth*, vol. 9, pp. 62–71.
- Ogilvy, R.D., Meldum, P.I. and Nissen, J., 1997. Quantitative mapping of waste coal for deposits by EM conductivity and VLF resistivity surveys. Extended Abstract, Proceedings, EEGS, Aarhus, pp. 45–47.
- Pati, P., Prakash, B., Awasthi, A.K., Acharya, V. and Singh, S., 2011. Concealed thrusts in the Middle Gangetic plain, India—a ground penetrating radar study proves the truth against the geomorphic features supporting normal faulting. *Journal of Asian Earth Sciences*, vol. 40, no. 1, pp. 315–325.
- Patidar, A.K., Maurya, M., Thakkar, M.G. and Chamyal, L.S., 2007. Fluvial geomorphology and neotectonic activity based on field and GPR data, Katrol hill range, Kachchh, Western India. *Quaternary International*, vol. 159, no. 1, pp. 74–92, DOI: 10.1016/j.quaint.2006.08.013.
- Pendias, A.K. and Pendias, H., 2001. *Trace Elements in Soil and Plants*. 3rd edn. Boca Raton, FL: CRC Press.
- Persson, K., 2005. Integrated geophysical-geochemical methods for archaeological prospecting. Licentiate thesis in Land and Water Resources Engineering. The Royal Institute of Technology, Stockholm.
- Prabhakar, V.N. and Korisetar, R., 2017. Ground survey to aerial survey: Methods and best practices in systematic archaeological explorations and excavations. *Current Science*, vol. 113, no. 10, pp. 1873–1890.
- Pridmore, D.F., Coggon, J.H., Esdale, D.J. and Lindemann, F.W., 1984. Geophysical exploration for nickel sulphide deposits in the Yilgarn Block, Western Australia. In: Buchanan, D.L., Jones, M.J. (eds), *Sulphide Deposits in Mafic and Ultramafic Rocks*. London: Institute of Mining and Metallurgy, pp. 22–34.
- Pujari, P.R., Pardhi, P., Muduli, P., Harkare, P. and Nanoti, M.V., 2007. Assessment of pollution near Landfill site in Nagpur, India by Resistivity Imaging and GPR. *Environment Monitoring and Assessment*, vol. 131, pp. 489–500.
- Radzevicius, S.J., Chen, C.C., Peters, L. and Daniels, J.J., 2003. Near-field dipole radiation dynamics through FDTD modelling. *Journal of Applied Geophysics*, vol. 52, pp. 75–91.
- Raiche, A., Annetts, D. and Sugeng, F., 2001. EM target response in complex hosts. 15th Geophysical Conference and Exhibition, Australian Society of Exploration Geophysicists, Abstracts, Brisbane.
- Raiche, A., Sugeng, F. and Wilson, G., 2007. Practical 3D EM inversion—P223F software suite. 19th Geophysical Conference and Exhibition, Australian Society of Exploration Geophysicists, Abstracts, Perth.
- Reddy, G.K., Menon, R., Seshunarayana, T. and Senthil Kumar, P., 2010. Soil-gas helium and surface waves detect fracture and fault zones in shallow granite basement. *Journal of Earth System Science*, vol. 119, pp. 745–751.
- Reynolds, J.M., 1997. *An introduction to applied and environmental geophysics*. Wiley, London.
- Reynolds, J.M., 1998. *An Introduction to Applied and Environmental Geophysics*. Chichester, New York: Wiley.
- Rout, S.P., Palanivel, K. and Kathiravan, R., 2018. Delineation of subsurface features of geological importance using GPR along coastal tract of Puri-Balasore districts of Odisha, India. *Journal of Applied Science and Computation*, vol. 5, no. 11, pp. 1081–1090.
- Safipour, R., Holz, S., Jegen, M. and Swidinsky, A., 2018. A first application of a marine inductive source EM configuration with remote electric dipole receivers: Palinuro Seamount Tyrrhenian Sea. *Geophysical Prospecting*, vol. 66, pp. 1415–1432.
- Sandmeier, K.J., 2016. *Reflexw Manual*. Karlsruhe, Germany: GPR's Instrument Manual.
- Sarris, A., Kalayci, T., Moffat, I. and Manataki, M., 2017. *An Introduction to Geophysical and Geochemical Methods in Digital Geoarchaeology*. Heidelberg, Germany: Springer, Digital Geoarchaeology, pp. 215–236.
- Sarvesh, K., Sanjit, K.P. and Arindam, G., 2020. Very low frequency electromagnetic (VLF-EM) study over Wajrakarur kimberlite Pipe 6 in Eastern Dharwar Craton, India. *Journal of Earth System Science*, vol. 129, pp. 2–10.

- Schlumberger, 1989. *Log Interpretation Principles/Applications*. Schlumberger Educational Services.
- Schmidt, A., 2013. *Earth Resistance for Archaeologists*. Lanham: AltaMira Press.
- Schroder, R., 1994. *Gültigkeitsbereich der Ebenen-Wellen-Theorie unter Berücksichtigung der Verschiebungsströme bezogen auf die geoelektrischen Meßmethoden VLF-R, (LF-R), CSAMT und Short WaveLoop-Loop, Diplomarbeit*. Institut für Geophysik und Meteorologie, Universität zu Köln. Köln, Germany: University of Cologne, p. 116.
- Scollar, I., 1962. Electromagnetic prospecting methods in archaeology. *Archaeometry*, vol. 5, pp. 146–153.
- Seidel, K., 2005. Elektromagnetische Zweispulen-Systeme, Prinzip der Methode. In: Knödel, K., Krummel, K., Lange, G. (eds), *Handbuch zur Erkundung des Untergrundes von Deponien und Altlasten, Bd 3 Geophysik*. Berlin, Heidelberg and New York: Springer, pp. 241–265.
- Seshunarayana, T., Satish Kumar, K., Deenanadh, S., Rama Mohan Rao, Y. and Senthil Kumar, P., 2008b. Subsurface imaging of dolerite dykes in Hyderabad granite terrain using multi-channel analysis of surface waves. In: Srivastava, R.K., Sivaji, C., Chalapathi Rao, N.V. (eds), *Indian Dykes*. New Delhi: Narosa Publishing House, pp. 331–338.
- Seshunarayana, T., Senthil Kumar, P., Sampath Rao, V., Mysaiah, D., Satish Kumar, K. and Pavan Kishore, P., 2008a. Multi-channel analysis of surface waves reveals shear zones and fractures in a shallow granite basement: Wailapally watershed near Hyderabad. *Journal Geological Society of India*, vol. 71, pp. 261–265.
- Shokri, N.H. and Amin, Z.M., 2016. Ground penetrating radar (GPR) application in archaeology: A case study. Research Gate, Conference Paper, IGCESH2016, Universiti Teknologi Malaysia, Johor Bahru, Malaysia, 15–17 August, p. 8.
- Shukla, S.B., Patidar, A.K. and Bhatt, N., 2008. Application of GPR in the study of shallow subsurface sedimentary architecture of Modwa spit, Gulf of Kachchh. *Journal of Earth System Science (India)*, vol. 117, no. 1, pp. 33–40.
- Sidorov, V.A. and Tickshaev, V.V., 1969. *Electrical Prospecting with Transient Field in Near Zone*. Saratov, Russia: Saratov University Press.
- Simon, F.X. and Moffat, I., 2015. Identification of shapes and uses of past landscapes through EMI survey. In: Sarris, A. (ed), *Best Practices of Geoinformatic Technologies for the Mapping of Archaeolandscapes*. Oxford: Archaeopress Publishing, pp. 25–34.
- Singh, K.K.K., 2003. Ground penetrating radar study for hydrogeological conditions related mining activity. *Environmental Geology*, vol. 44, pp. 20–27.
- Smith, J.T. and Booker, J.R., 1991. Rapid inversion of two- and three-dimensional magnetotelluric data. *Journal of Geophysical Research*, vol. 96, pp. 3905–3922.
- Soni, K., Singh, K.K.K., Prakash, A. and Singh, K.B., 2007. Shallow cover over coal mining: A case study of subsidence at Kamptee Colliery, Nagpur, India. *Bulletin of Engineering Geology and Environment*, vol. 66, no. 3, pp. 311–318.
- Sonkamble, S., Chandra, S., Nagaiah, E., Dar, F.A., Somvanshi, V.K. and Ahmed, S., 2013. Geophysical signatures resolving hydrogeological complexities over hard rock terrain—a study from Southern India. *Arabian Journal of Geosciences*, vol. 7, pp. 2249–2256.
- Soon, Y., Gan, K.B. and Abdullah, M., 2015. Development of Very Low Frequency (VLF) data acquisition system using Raspberry Pi. International Conference on Space Science and Communication (Icon Space), DOI: 10.1109/ICONSPACE.2015.7283807 Corpus ID: 43841376.
- Spies, B., 1980. Recent developments in the use of surface electrical methods for oil and gas exploration in the Soviet Union. *Geophysics*, vol. 48, pp. 1102–1112.
- Sravanthi, S., Malik, J.N. and Vikrama, B., 2012. Ground penetrating radar investigations at Ahichchhatra: An attempt to identify buried subsurface structures. 14th International Conference on Ground Penetrating Radar (GPR), Shanghai, China, 4–8 June, pp. 625–630.
- Strack, K.-M., 1992. *Exploration with Deep Transient Electromagnetics*. Amsterdam: Elsevier, p. 373.
- Strack, K.-M., Hanstein, T., LeBrocq, K., Moss, D.C., Vozoff, K. and Wolfgram, P.A., 1989. Case histories of LOTEM surveys in hydrocarbon prospective areas. *Article in First Break (Research Gate)*, vol. 7, no. 12, p. 468.
- Streich, R., van der Kruk, J. and Green, A.G., 2006. Three dimensional multicomponent georadar imaging of sedimentary structures. *Near Surface Geophysics*, vol. 4, pp. 4–39.
- Supper, R., Motschka, K., Ahl, A., Bauer-Gottwein, P., Gondwe, B., Alonso, G.M., Roemer, A., Ottowitz, D. and Kinzelbach, W., 2009. Spatial mapping of submerged cave systems by means of airborne electromagnetics: An emerging technology to support protection of endangered karst aquifers. *Near Surface Geophysics*, vol. 7, no. 5–6, pp. 613–627.

- Swain, A.K., 2010: A report on the participation in the first Indian Scientific Expedition to South Pole. Unpublished Report, Antarctica Division, Geological Survey of India, Faridabad, pp. 1–15.
- Swain, A.K., 2018. Deep probing through ground penetrating radar (GPR)—case study from Antarctica. *Journal of Engineering Geology*, vol. 43, no. 1 & 2, pp. 38–47.
- Swain, A.K. and Goswami, S., 2014. Continuous GPR survey using multiple low frequency antennas—case studies from Schirmacher Oasis, East Antarctica. *International Journal of Earth Science and Engineering*, vol. 7, no. 4, pp. 1623–1629.
- Tabbagh, A., 1986. Applications and advantages of the Slingram EM method for archaeological prospecting. *Geophysics*, vol. 51, pp. 576–584.
- Tao, W., Xiong, Z., Deng Xi, X. and Xu, Y., 2013. The investigations of South Atlantic Ridge 13.2°S hydrothermal area. *Acta Oceanologica Sinica*, vol. 32, no. 12, pp. 68–74.
- Teixido, T., Peña, J.A., Lopez Castro, J.L. and Adrover, A., 2010. Prospección Geofísica en varias zonas del yacimiento arqueológico de Utica, Túnez. Universidad de Granada- OTRI-IAG; technical report ref. AGA-78.
- Tezkan, B., 1999. A review of environmental applications of quasistationary electromagnetic techniques, Surveys. *Geophysics*, vol. 20, pp. 279–308.
- Tezkan, B., Goldman, M., Greinwald, S., Hordt, A., Muller, I., Neubauer, F.M. and Zacher, G., 1996. A joint application of radio magnetotellurics and transient electromagnetics to the investigation of a waste deposit in Cologne (Germany). *Journal of Applied Geophysics*, vol. 34, pp. 199–212.
- Tite, M. and Mullins, C., 1969. Electromagnetic prospecting—a preliminary investigation. *Prospezioni Archeologiche*, vol. 4, pp. 95–102.
- Tite, M. and Mullins, C., 1970. Electromagnetic prospecting on archaeological sites using a soil conductivity meter. *Archaeometry*, vol. 12, pp. 97–104.
- Trench, A. and Williams, P.K., 2019. Nickel: Application of geophysics to nickel sulphide exploration in the Kambalda district, Western Australia. *ASEG Extended Abstracts*, 1994, vol. 1, pp. 169–180, DOI: 10.1071/ASEGSpec07_10.
- Tronicke, J., Dietrich, P., Wahlig, U. and Appel, E., 2002. Integrating surface georadar and cross-hole radar tomography: A validation experiment in braided stream deposits. *Geophysics*, vol. 78, pp. 1516–1523.
- Turberg, P., Muller, I. and Flury, F., 1994. Hydrogeological investigation of porous environments by radio-magnetotelluric-resistivity (RMT-R, 12–240 kHz). *Journal of Applied Geophysics*, vol. 31, pp. 133–143.
- Ulriksen, C., 1982. Application of the impulse radar to civil engineering. Ph.D. thesis. Lund University of Technology, Sweden, p. 179.
- Vanyan, L.L., Bobrovnikov, L.Z. and Loshenitzina, V.L., 1967a. *Electromagnetic Depth Soundings*. Moscow: Consultants Bureau.
- Vanyan, L.L., Bobrovnikov, L.Z. and Loshenitzina, V.L., 1997b. *Electromagnetic Soundings* (in Russian). Berlin, Germany: Scientific Word.
- Vaughan, C., 1986. Ground penetrating radar surveys used in archaeological investigations. *Geophysics*, vol. 51, pp. 595–604.
- Vogelsang, D., 1995. *Environmental Geophysics: A Practical Guide*. Berlin, Germany: Springer Verlag.
- Von Hippel, A.R., 1954a. *Dielectrics and Waves*. Cambridge, MA: MIT Press.
- Von Hippel, A.R. (ed), 1954b. *Dielectric Materials and Applications*. Cambridge, MA: MIT Press.
- Vonder Muhll, D.S. and Holub, P., 1992. Borehole logging in Alpine permafrost, Upper Engadine, Swiss Alps. *Permafrost and Periglacial Processes*, vol. 3, no. 2, pp. 125–132.
- Vonder Muhll, D.S., Noetzi, J., Roer, I., Makowski, K., Delaloye, R., Gruber, S., Hauck, C., Hoelzle, M., Kääh, A., Luetscher, M., Phillips, M., Salzmann, N., Stucki, T. and Haeberle, W., 2005. Permafrost der Schweizer Alpen 2002/03 und 2003/04. *Die Alpen*, vol. 81, no. 10, pp. 2431. DOI: 10.13140/RG.2.1.1302.3440.
- Voxler Team, 2015. *Voxler Full User Guide Manual. Instrumentation Manual*, United States: Golden, Colorado.
- Vozoff, K., 1991. The magnetotelluric method. In: Nabighian, M.N. (ed), *Electromagnetic Methods in Applied Geophysics 3*. SEG, pp. 641–711.
- Wait, J.R., 1982. *Geo-Electromagnetism*. New York: Academic Press Inc.
- Wang, Z., Wan, B. and Han, M., 2020. A three-dimensional visualization framework for underground geohazard recognition on urban road-facing GPR data. *ISPRS International Journal of Geo-Information*, vol. 9, DOI: 10.3390/ijgi9110668.
- Weidelt, P., 1997. Grundlagen der Geoelektrik. In: Knodel, K., Krummel, H., Lange, G. (eds), *Handbuch zur Erkundung des Untergrundes von Deponien und Altlasten, Band3, Geophysik*. Heidelberg, Germany: Springer-Verlag, pp. 65–83.

- Williamson, P.R., 1991. A guide to the limits of resolution imposed by scattering in ray tomography. *Geophysics*, vol. 56, pp. 202–207.
- Williamson, P.R. and Worthington, M.H., 1993. Resolution limits in ray tomography due to wave behaviour: Numerical experiments. *Geophysics*, vol. 58, pp. 727–735.
- Wilsom, J., 1989. *Field Geophysics*. Hoboken, NJ: Wiley.
- Wilson, G.A., Cox, L.H. and Zhdanov, M.S., 2010. Practical 3D inversion of entire airborne electromagnetic surveys. *Preview*, vol. 146, pp. 29–33.
- Witherly, K., 2000. The quest for the Holy Grail in mining geophysics: A review of the development and application of airborne EM systems over the last 50 years. *The Leading Edge*, vol. 19, pp. 270–274, DOI: 10.1190/1.1438586.
- Wolfgram, P. and Golden, H., 2001. Airborne EM applied to sulphide nickel—examples and analysis. *Exploration Geophysics*, vol. 32, no. 4, pp. 136–140, DOI: 10.1071/EG01136.
- Wright, D., Ziolkowski, A. and Hobbs, B., 2002. Hydrocarbon detection and monitoring with a multicomponent transient electromagnetic (MTM) survey. *The Leading Edge*, vol. 21, pp. 852–864.
- Wu, T.T. and King, R.W.P., 1965. The cylindrical antenna with non-reflecting resistive loading. *IEEE Trans Antennas and Propagation AP-13*, United States: Piscataway, New Jersey, pp. 369–373. DOI:10.1109/TAP.1965.1138429 Corpus ID: 121890808.
- Wynn, J. and Sherwood, S., 1984. The self-potential (SP) method—an inexpensive reconnaissance and archaeological mapping tool. *Journal of Field Archaeology*, vol. 11, pp. 195–204.
- Xue, G.Q., Qin, K.Z., Li, X., Li, G.M., Qi, Z.P. and Zhou, N.N., 2012. Discovery of the large-scale porphyry molybdenum deposit in Tibet through modified TEM exploration method. *Journal of Environmental & Engineering Geophysics*, vol. 17, no. 1, pp. 19–25.
- Zacher, G., 1992. Hochfrequente Elektromagnetik auf Altlasten, DGG-Kolloquium Elektromagnetische Tiefenforschung Borkheide, G.F.Z. Potsdam, pp. 357–363.
- Zacher, G., Tezkan, B., Neubauer, F.M., Hordt, A. and Muller, I., 1996. Radio magnetotellurics: A powerful tool for waste-site exploration. *European Journal of Environmental and Engineering Geophysics*, vol. 1, pp. 139–159.
- Zhang, D., Gong, H., Li, J., Wu, Z. and Liu, S., 2022. Integrated ground penetrating radar and DGPS method for the continuous and longdistance GPR survey in the rugged terrain. *Acta Geophysica*, vol. 70, pp. 537–546.
- Zhang, Y. and Ni, P., 2018. Design optimization of room and pillar mines: A case study of the Xianglushan tungsten mine Q. *Journal of Engineering Geology and Hydrogeology*, vol. 51, no. 3, pp. 352–364.
- Zhdanov, M.S., 2010. Electromagnetic geophysics: Notes from the past and the road ahead. *Geophysics*, vol. 75, no. 5, pp. 75A49–75A66, 11 FIGS, DOI: 10.1190/1.3483901.
- Zhdanov, M.S., 2018. In *Foundations of Geophysical Electromagnetic Theory and Methods*. 2nd edn. Amsterdam, Netherlands: Elsevier.
- Zhdanov, M.S. and Keller, G.V., 1994. *The Geoelectrical Methods in Geophysical Exploration*. Amsterdam, Netherlands: Elsevier.
- Zhu, X.Y., Wand, J.B., Wang, Y.L. and Chen, X.Y., 2015. The role of magma-hydrothermal transition fluid in the skarn-type tungsten mineralization process: A case study from the Shizhuyuan tungsten and tin polymetallic ore deposit. *Acta Petrologica Sinica*, vol. 31, no. 3, pp. 891–905 (in Chinese with English abstract).
- Ziolkowski, A.M., Hobbs, B.A. and Wright, D.A., 2007. Multitransient electromagnetic demonstration survey in France. *Geophysics*, vol. 72, no. 4, pp. F197–F209, DOI: 10.1190/1.2735802.

9 Magnetotelluric (MT) Method

The magnetotelluric (MT) method is a passive electromagnetic (EM) technique for which the electric field ($\vec{E}$) and the magnetic field ($\vec{B}$ or $\vec{H}$) are measured in orthogonal directions on the Earth's surface. Investigation depth ranges from 300 m belowground by recording higher frequencies down to 10,000 m or deeper with low-frequency or long-period soundings. Since the year 1940, the methodology of MT geophysical survey has been in use in Japan, and later, it became an important geophysical method of exploration for France and the USSR during the early 1950s. Indian geoscientific research and academic institutes started to deploy MT surveys for deeper subsurface investigations in association with the seismic method since the year 1985. MT geophysical survey has now become an international academic discipline and is being used in the exploration of the Earth all over the globe. The MT theory is based on the telegraphic equation of electromagnetic waves, in which the use of uncontrollable telluric current leads to changes in electrical and magnetic fields. Five orthogonally oriented, electric field (E_x ; E_y), and magnetic field (H_x ; H_y ; H_z) sensors record the respective changes in both electric and magnetic fields. The ratio of $\frac{E_x}{H_y}$ gives the wave impedance, and using the Cagniard equation, the apparent resistivity is estimated for the given frequency of the site. The physical parameters measured are to be analyzed in detail along with appropriate data inversion procedures that allow the determination of the resistivity distribution in the subsurface, on depth scales ranging from a few tens of metres to hundreds of kilometres (Tikhonov, 1950; Cagniard, 1953). The MT method is passive in the sense that it utilizes naturally occurring geomagnetic variations as a power source. The periodicity of the source, as well as the resistivity distribution of the subsurface, has influence on the depth of information for which the user is interested.

9.1 SOURCE OF SIGNAL

Earth's natural electromagnetic (EM) field originates due to the interaction of the solar flares with the Earth's magnetic field in the magnetosphere. The solar flares are of irregular nature, and solar emissions continuously vary with time. Solar emissions are minimum during solar quiet days, and the emission increases during a solar disturbance. X-rays, γ -rays, electrons, and protons in the form of solar plasma, cosmic rays, and ultraviolet rays are emitted continuously from the emission surface. X-rays and γ -rays, being high-energy and highly penetrating radiations, enter into the atmospheric level of the Earth and ionize the ionosphere and Van Allen radiation belts. Ultraviolet rays (UV) penetrate up to the ozonosphere level, and most of these rays get absorbed there. Electrons and protons in the form of solar plasma interact with the magnetic lines of forces in the magnetosphere and get trapped. They spiral around the magnetic lines of forces and start traveling towards the geomagnetic latitudes of the Earth. The geomagnetic fluctuations range between periods of 10^{-3} s and 10^5 s (or between frequencies of 10^3 Hz and 10^{-5} Hz), depending on their origin. Since the number of electrons and protons getting trapped at a particular moment is highly variable, they generate time-varying EM fields (Wameyo, 2005, KenGen, p. 25).

This time-varying EM field is superimposed on the stronger, stable magnetic field of the Earth, which is originated due to the dynamo current in the Earth's core. This time-varying EM field generates different time-varying events known as geoelectric and geomagnetic pulsations and micro-pulsations, with periods of up to 600 s, as shown in Table 9.1. Longer periods can be divided into variations from quiet and disturbed days. Quiet days comprise, for example, the solar quiet (*Sq*) and lunar (*L*) variations and the solar flare effect (*sfe*). However, on disturbed days, sudden storm commencements (SSC) cause an increase in ring currents and bay anomalies, which result in

TABLE 9.1**Typical Classification Scheme for Geomagnetic Pulsation Utilized in the MT Method**

	Continuous Pulsations					Irregular Pulsations	
	Pc1	Pc2	Pc3	Pc4	Pc5	Pi1	Pi2
<i>T</i> in s	0.2–5	5–10	10–45	45–150	150–600	1–40	40–150
<i>F</i> in Hz	0.2–5	0.1–0.2	0.022–0.100	0.007–0.022	0.002–0.007	0.025–1	0.002–0.025

higher-amplitude variations than the *Sq* (Campbell, 1997). Since solar emissions have a very wide frequency spectrum, geophysicists use these signals to explore the interior structure of the Earth.

The wide frequency spectrum of the MT field cannot be fully explained on the basis of these pulsations and micropulsations. The high-frequency components of the magnetic field, where the frequencies are of the order of 1 Hz to several kilo Hz, originate due to thunderstorm activities in between the Earth's surface and the ionosphere. The signals travelling within the waveguide are also known as spherics. Between 0.5 and 5 Hz lies the dead band, at which the natural EM fluctuations have a low intensity. MT measurements in this frequency range usually suffer from poor data quality.

It may be noted that the longer the periods, the higher the energy levels of the signals, and the higher the resistivity of the Earth's crust, the greater the depth of penetration of the signal. These properties make the EM depth sound effective for the study of the Earth's interior. Almost all plane or nearly plane EM waves originate in the magnetosphere and propagate towards the surface of the Earth. Subsequently, they diffuse through the Earth's crust and penetrate deep inside. It is clear from the preceding discussions that for greater penetration of the EM signals, geophysicists search granite/granodiorite windows which have lower conductivities.

9.1.1 THEORETICAL CONSIDERATIONS

The magnetotelluric (MT) method is one of the geophysical exploration techniques in which the conductivity structure of the Earth is studied utilizing the naturally existing electromagnetic fields of the Earth. MT surveys are mostly deployed for probing greater depths of the Earth (tens of kilometres or even more) using very-low-frequency MT signals (10^{-2} Hz to 10^{-5} Hz). From such surveys, resistivity distribution along depth is estimated. This leads to the mapping of (1) electrical conductivity of the Earth's lithosphere, (2) electrical conductivity of geothermal areas and magma chambers, and (3) lineaments/fractures/faults. Detailed MT studies are being carried out in different countries to explore the possibilities of detection of the (1) olivine–spinel transition zone, (3) Mesozoic sediments below flood basalts, (3) subduction zone, and (4) conducting zones. The audio frequency magnetotelluric (AMT) method is often used for mapping metallic mineralized zones and conductivity contrasts at shallow depth.

The time-varying electromagnetic field is recorded in five channels, viz., E_x, E_y, H_x, H_y and H_z , which are measured in the form of time series in time domain. These signals pass through a series of data processing software, viz., moving average algorithm, detrending algorithm, filters, windows (cosine window, Hanning and Hamming windows, Parzen window, etc.), convolution, autocorrelation, cross-correlation, fast Fourier transform (FFT), etc. Autopower spectra and cross-power spectra are obtained from these time series analyses. Once the electric and magnetic fields are obtained in the frequency domain, they go through the rigorous checking by monovariate and bivariate coherency gates. The computation of MT impedance tensors starts using the formulations given by Sims, 1971, Kao and Rankin (1977), and Gamble et al. (1979). At this stage, we get all the components of the MT impedance tensors, like Z_{xx}, Z_{xy}, Z_{yx} , and Z_{yy} . Cagniard impedances (Cagniard, 1953) are computed using the off-diagonal elements. After mathematical rotation

(Swift, 1967; Vozoff, 1991), TE (transverse electric) and TM (transverse magnetic) MT tensors are estimated. These four elements of the impedance tensor describe ellipses of the same size and ellipticity. However, the sizes of the ellipses and their ellipticities vary with the periods of MT signals.

Spectral power distribution is done by taking a number of samples equal to a power of 2 (from 256 to 8,192). Then an FFT algorithm is applied, which yields from 256 to 4,096 complex spectral values corresponding to different frequencies. Each spectral value is defined by an in-phase and an out-of-phase component. The next stage will be to correct the computed values for the frequency response of the sensor (amplitude and phase) as well as for the gain used during recording.

At a given frequency (ω), there are generally the following relations between the five spectral components H_x, H_y, H_z, E_x, E_y .

$$E_x(\omega) = Z_{xx}(\omega) H_x(\omega) + Z_{xy}(\omega) H_y(\omega)$$

$$E_y(\omega) = Z_{yx}(\omega) H_x(\omega) + Z_{yy}(\omega) H_y(\omega)$$

$$H_z(\omega) = Z_{zx}(\omega) H_x(\omega) + Z_{zy}(\omega) H_y(\omega)$$

where $Z_{xx} = \frac{E_x}{H_x}$, $Z_{xy} = \frac{E_x}{H_y}$, $Z_{yx} = \frac{E_y}{H_x}$, $Z_{yy} = \frac{E_y}{H_y}$; Z_{xx} and Z_{yy} are diagonal elements, while Z_{xy} and Z_{yx} are off-diagonal elements. One intends to compute the complex impedances Z_{xx} , Z_{xy} , Z_{yx} , Z_{yy} , Z_{zx} , and Z_{zy} .

Calculations are done in the following procedure, for each frequency window. One computes the average autocorrelation function for each spectral component, for example, $H_x H_x^*$, $E_x E_x^*$, $H_y H_y^*$, $E_y E_y^*$, $H_z H_z^*$. One computes the average intercorrelation for each two of these components, for example, $E_x H_y^*$, $E_y H_x^*$, etc. The signals are first allowed to pass through the coherency gates, and then the monovariate and bivariate coherencies are examined, which reflect the measurement quality. Good-quality data yield coherency comprises between 0.95 and 1.00. Monovariate coherency is given by

$$\text{Coh}(E_x H_y) = \frac{\langle E_x H_y \rangle}{\left[\langle E_x E_x^* \rangle \langle H_y H_y^* \rangle \right]^{\frac{1}{2}}}$$

Bivariate coherency is given by

$$\text{Coh}(E_i^{\text{Pred}} E_i) = \frac{\langle E_i^{\text{Pred}} E_i \rangle}{\left[\langle E_i^{\text{Pred}} E_i^{\text{Pred}*} \rangle \langle E_i E_i^* \rangle \right]^{\frac{1}{2}}}$$

These coherencies are expressed in terms of auto- and cross-powers of the E and H fields. Here, $\langle E_x H_y \rangle$ is the cross-power spectrum of E_x and H_y , and $\langle E_x E_x^* \rangle$, $\langle H_y H_y^* \rangle$ are the autopower spectrum of E_x and H_y , whereas E_x^* and H_y^* are the complex conjugates of E_x and H_y . E_i^{pred} is the predicted E field obtained from the measured magnetic fields.

9.2 BEHAVIOUR OF ELECTROMAGNETIC FIELDS

The propagation of EM fields can be described by a set of four relations, called the Maxwell equations of electromagnetic wave. All entities (E, D, H, B, J) of electrical and magnetic fields that define the electromagnetic wave are expressible in terms of vector. Given the following, a polarizable and magnetizable medium containing no electric and magnetic sources, the following relationships hold at all times for all frequencies (Feynman et al., 2005):

$$\nabla \times \mathbf{E} = -\frac{\partial \mathbf{B}}{\partial t}, \text{ Faraday's law} \quad (9.1a)$$

Faraday's law in eq. (9.1a) shows the coupling of an induced electric field E (V/m) in a closed loop due to a time-varying magnetic field B (tesla; T) along the axis of the induced electric field.

$$\nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t}, \text{ Ampere's law} \quad (9.1b)$$

Curls of magnetic fields with a magnetic intensity H (A/m) are caused by electric current densities J (A/m²) and time-varying electric displacements.

$$\nabla \cdot \mathbf{B} = 0, \text{ Gauss's law for magnetism} \quad (9.1c)$$

Eq. 9.1c states that magnetic fields, defined by the magnetic induction B (in tesla), are always source-free, and no free magnetic poles exist.

$$\nabla \cdot \mathbf{D} = \rho, \text{ Gauss's law} \quad (9.1d)$$

The electric displacement D (C/m²) is solely due to an electric charge density.

$$\nabla \cdot \mathbf{J} + \frac{\partial \rho}{\partial t} = 0, \text{ equation of continuity} \quad (9.2)$$

The expressions for Faraday's and Ampere's law are initially decoupled. However, in the presence of a linear, isotropic medium, the material equations can be introduced:

$$\mathbf{B} = \mu \mathbf{H} \quad (9.3a)$$

$$\mathbf{D} = \epsilon \mathbf{E} \quad (9.3b)$$

Furthermore, currents in the presence of an electric field can only flow if the media has a conductivity σ (in S/m) other than zero (Ohm's law):

$$\mathbf{J} = \sigma \mathbf{E} \quad (9.4)$$

μ = magnetic permeability = $\mu_0 \mu_r$

μ_0 = magnetic permeability in free space

μ_r = relative magnetic permeability

ϵ = dielectric permittivity = $\epsilon_0 \epsilon_r$

ϵ_0 = dielectric permittivity of free space

ϵ_r = relative dielectric permittivity

$\mu_0 = 4\pi \times 10^{-7}$ H/m = magnetic permeability of free space

$\epsilon_0 = 8.85 \times 10^{-12}$ F/m = dielectric permittivity of free space

Variations in μ_r and ϵ_r are assumed negligible compared to variations in the bulk conductivity σ of rocks. The variables μ , ϵ , and σ are the properties depending entirely on the material through which the electromagnetic waves propagate. For homogeneous materials, ϵ and μ are constants, while for inhomogeneous materials, they depend on time and position within the material. In the case of isotropic material, μ and ϵ are scalars, and for anisotropic materials, they are represented by tensors. The variation in the resistivity of the Earth is highly dependent on parameters that characterize the reservoir. These variations are related to the properties of interests and depend on porosity of the rock, alteration, water saturation, salinity of the water, temperature, etc.

9.2.1 ASSUMPTIONS OF MAGNETOTELLURICS

The following are the considerable assumptions applicable in electromagnetic induction in the Earth:

- The Earth does not generate electromagnetic (EM) energy but only dissipates or absorbs it.
- Maxwell's electromagnetic (EM) equations are obeyed.
- All electromagnetic fields are treated as conservative and analytic away from their sources.
- The passive electromagnetic source fields, being generated by large-scale ionospheric current systems that are relatively far away from the Earth surface, may be treated as uniform, plane-polarized electromagnetic waves impinging on the Earth at near-vertical incidence.
- Accumulation of free charges cannot be expected to be sustained within a layered Earth. However, in multidimensional Earth, charges can accumulate along discontinuities. The Earth behaves as an ohmic conductor and obeying the equation $\mathbf{J} = \sigma \mathbf{E}$, where $\mathbf{J}$ is total electric current density (in A), σ is the conductivity of the medium (in simon per metre, Sm^{-1}), and $\mathbf{E}$ is the electric field (in volt per metre, Vm^{-1}).
- The time-varying displacement currents (arising from polarization effects) are negligible compared with time-varying conduction currents and promote the treatment of electromagnetic induction in the Earth purely as a diffusion process.
- The variations in the magnetic permeabilities and electrical permittivities of rocks are assumed negligible.

9.3 ELECTROMAGNETIC INDUCTION IN A HOMOGENEOUS HALF-SPACE

In the case of a homogeneous and isotropic medium, substituting eqs. 9.3 and 9.4 into eqs. 9.1 and assuming the harmonic dependence of the oscillating EM fields, $e^{i\omega t}$, leads to the following relationships:

$$\nabla \times \mathbf{E} = -i\omega\mu\mathbf{H} \quad (9.5)$$

$$\nabla \times \mathbf{H} = (\sigma + i\omega\epsilon)\mathbf{E} \quad (9.6)$$

$\omega = 2\pi f$ = angular frequency

$f = \frac{1}{T}$ = frequency of oscillation

T = period of oscillation

$i = \sqrt{-1}$ = imaginary unit of complex number

There is no charge accumulation in a homogeneous conductive medium; therefore, Gauss's law implies that $\nabla \cdot \mathbf{D} = \nabla \cdot (\epsilon \mathbf{E}) = 0$. Considering the EM waves to be vertically incident plane waves only, it follows that $\nabla_x = \nabla_y = 0$, and the EM field has only a z component. Using the definition of curl, eqs. 9.5 and 9.6 give:

$$\frac{\partial E_z}{\partial y} - \frac{\partial E_y}{\partial z} = -i\omega\mu H_x \quad (9.7a)$$

$$\frac{\partial E_x}{\partial z} - \frac{\partial E_z}{\partial x} = -i\omega\mu H_y \quad (9.7b)$$

$$\frac{\partial E_y}{\partial x} - \frac{\partial E_x}{\partial y} = -i\omega\mu H_z \quad (9.7c)$$

$$\frac{\partial H_z}{\partial y} - \frac{\partial H_y}{\partial z} = (\sigma + i\omega\epsilon) E_x \quad (9.8a)$$

$$\frac{\partial H_x}{\partial z} - \frac{\partial H_z}{\partial x} = (\sigma + i\omega\epsilon) E_y \quad (9.8b)$$

$$\frac{\partial H_y}{\partial x} - \frac{\partial H_x}{\partial y} = (\sigma + i\omega\epsilon) E_z \quad (9.8c)$$

Since the H field propagates vertically downward in the z direction, while the induced E field does not have a z component ($E_z = 0$), the plane wave assumption states that the inducing magnetic field only has horizontal components due to large the distance to the source and that it is therefore valid for periods up to 10^5 s. This also implies that the H_z component is equal to 0. Therefore,

$\frac{\partial E_x}{\partial y}$ and $\frac{\partial E_y}{\partial x}$ have to be zero too. Eq. 9.7 and 9.8 become:

$$\frac{\partial^2 E_x}{\partial z^2} = -i\omega\mu \frac{\partial H_y}{\partial z} = i\omega\mu(\sigma + i\omega\varepsilon)E_x \quad (9.9a)$$

$$\frac{\partial^2 E_y}{\partial z^2} = i\omega\mu \frac{\partial H_x}{\partial z} = i\omega\mu(\sigma + i\omega\varepsilon)E_y \quad (9.9b)$$

Following a similar approach, we find a relation for H_x and H_y :

$$\frac{\partial^2 H_x}{\partial z^2} = i\omega\mu \frac{\partial E_y}{\partial z} = i\omega\mu(\sigma + i\omega\varepsilon)H_x \quad (9.10a)$$

$$\frac{\partial^2 H_y}{\partial z^2} = i\omega\mu \frac{\partial E_x}{\partial z} = i\omega\mu(\sigma + i\omega\varepsilon)H_y \quad (9.10b)$$

Eq. 9.9 and 9.10 may be expressed as (Helmholtz wave equation in the frequency domain)

$$\frac{\partial^2 E_{x,y}}{\partial z^2} = k^2 E_{x,y} \quad (9.11)$$

$$\frac{\partial^2 H_{x,y}}{\partial z^2} = k^2 H_{x,y} \quad (9.12)$$

where $k^2 = i\omega\mu\sigma - \omega^2\mu\varepsilon$ = wave number of the EM wave. In a conductive medium, $\sigma \gg \omega\varepsilon$, the wave number k represents the quasistatic approximation, $k \approx \sqrt{i\omega\mu\sigma}$. In the wave equation, the speed of the wave is described as inverse square root of the coefficient of the time derivative; then, both fields have the same speed of $c = \frac{1}{\sqrt{\mu\varepsilon}} = 3 \times 10^8$ m/s, the speed of light. The preceding results completely describe the electric and magnetic fields in a homogeneous medium. A general solution to eqs. 9.11 and 9.12 is

$$E_{x,y}(t) = (A_{x,y}e^{kz} + B_{x,y}e^{-kz})e^{i\omega t} \quad (9.13)$$

$$H_x(t) = \frac{k}{i\omega\mu} (A_y e^{kz} - B_y e^{-kz}) e^{i\omega t} \quad (9.14)$$

$$H_y(t) = \frac{-k}{i\omega\mu} (A_x e^{kz} - B_x e^{-kz}) e^{i\omega t} \quad (9.15)$$

where $A_{x,y}$ and $B_{x,y}$ are constants to be determined from the boundary conditions. Since the EM field in MT measurements originates from the ionosphere, it applies that all the quantities must remain finite when z goes to infinity. Thus, the first terms in eqs. 9.13, 9.14, and 9.15 vanish:

$$E_{x,y}(t) = B_{x,y} e^{-kz} e^{i\omega t} \quad (9.16)$$

$$H_x(t) = \frac{1}{i\omega\mu} \frac{\partial E_y}{\partial z} = \frac{-k}{i\omega\mu} B_y e^{-kz} e^{i\omega t} \quad (9.17)$$

$$H_y(t) = \frac{-1}{i\omega\mu} \frac{\partial E_x}{\partial z} = \frac{k}{i\omega\mu} B_x e^{-kz} e^{i\omega t} \quad (9.18)$$

Eqs. 9.16, 9.17, and 9.18 illustrate the superposition of two characteristics of the field. The fields vary sinusoidally with depth due to the $e^{i\omega t}$ term but also show a depth-dependent attenuation due to e^{-kz} term.

The ratio between the orthogonal components of the electric and magnetic field intensity is known as the characteristic impedance, Z , and represents a characteristic measure of the EM properties of the subsurface. It has units of resistance (Ω). For a uniform Earth, the surface impedance is related to the true resistivity of the rock. The ratio between the orthogonal component of electric and magnetic field yields the Tikhonov–Cagniard impedance:

$$Z_{xy} = \frac{E_x}{H_y} = \frac{i\omega\mu}{k} = \frac{i\omega\mu}{\sqrt{i\omega\mu\sigma}} \approx \sqrt{\frac{\omega\mu}{\sigma}} e^{\frac{i\pi}{4}} \quad (9.19a)$$

$$Z_{yx} = \frac{E_y}{H_x} = \frac{-i\omega\mu}{k} = -Z_{xy} \quad (9.19b)$$

The dimensions of the electric and magnetic field intensity are V/m and A/m, respectively, but it is common to measure these properties in mV/km. The entire eq. 9.19 has a constant phase of $\varphi = \frac{\pi}{4}$, and the magnetic field lags behind the electric field by 45° for a homogeneous Earth. The phase φ is independent of resistivity. According to Cagniard (1953), the resistivity of the half-space in units of $\Omega\cdot m$, by reordering eq. 9.19, gives

$$\rho = \frac{1}{\omega\mu} |Z|^2 = \frac{T}{2\pi\mu} |Z|^2 = 0.2T \frac{|E_x|^2}{|H_y|^2} \quad (9.20)$$

$$Z = \frac{E}{H} \text{ and } Z_{xy} = \frac{E_x}{H_y}, \quad Z_{yx} = \frac{E_y}{H_x} \quad (9.21)$$

When the Earth resistivity is nonuniform, the right-hand sides of eq. 9.20 provide apparent resistivities (instead of true resistivity), ρ_a , which are frequency (period) dependent, as is the case with 1-D, 2-D, or 3-D situations. In the frequency domain, $Z(\omega)$ is complex and has an associated phase (φ). So phase of Z_{xy}

$$\varphi = \tan^{-1} \frac{\text{Im} \left| \frac{E_x}{H_y} \right|}{\text{Re} \left| \frac{E_x}{H_y} \right|} \quad (9.22)$$

The impedance phase gives additional information about the conductive structure. For a nonhomogeneous medium, phase $\neq \frac{\pi}{4}$.

9.3.1 SKIN DEPTH

The wave number $k = \sqrt{i} \sqrt{i\omega\mu\sigma} = \sqrt{\omega\mu\sigma} e^{\frac{i\pi}{4}} = (1 + i) \sqrt{\frac{\omega\mu\sigma}{2}}$. The inverse of the real part of k is also known as the skin depth or penetration depth (δ).

$$\delta = \frac{1}{\text{Re}k} = \sqrt{\frac{2}{\omega\mu\sigma}} \quad (9.23)$$

Since the magnetic permeability μ does not vary substantially in the Earth, eq. 9.23 can be approximated as a formula for the skin depth (in m):

$$\delta \text{ (T)} = 504 \sqrt{T\rho_a} \quad (9.24)$$

where $\rho_a = \frac{1}{\sigma_a}$ is the apparent resistivity or the equivalent average resistivity of the uniform half-space (in $\Omega\text{-m}$). EM fields are attenuated to a value of e^{-1} of their surface amplitude at a depth

$\delta \text{ (T)}$ (Zhang et al., 1993). This relation is crucial, as eq. 9.24 states that the skin depth is only due to the bulk conductivity of the overlying material and the period range used.

9.3.2 ANISOTROPIC HALF-SPACE

In an anisotropic media, the conductivity (σ) is represented by a second-rank 3-D tensor (Heise et al., 2006). Ohm's law can be expressed in 3-D form (see eq. 9.4 for the isotropic case):

$$\begin{pmatrix} j_x \\ j_y \\ j_z \end{pmatrix} = \begin{pmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz} \end{pmatrix} \begin{pmatrix} E_x \\ E_y \\ E_z \end{pmatrix}$$

where σ is symmetric ($\sigma_{ij} = \sigma_{ji}$) and non-negative-definite (Weidelt, 1999). With these properties, the tensor can be diagonalized and expressed by three principal conductivities, σ_1 , σ_2 , σ_3 , and three rotation angles that relate the orientation of the tensor's principal axes (x' , y' , z') to the reference frame (x , y , z). Following the notation of Pek and Santos (2006), the rotation angles are referred to as Euler angles α_L , α_D , α_S :

$$\sigma = \begin{pmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz} \end{pmatrix} = R_z(-\alpha_S) R_x(-\alpha_D) R_z(-\alpha_L) \underbrace{\begin{pmatrix} \sigma_1 & 0 & 0 \\ 0 & \sigma_2 & 0 \\ 0 & 0 & \sigma_3 \end{pmatrix}}_{\sigma(x', y', z')} R_z(\alpha_L) R_x(\alpha_D) R_z(\alpha_S)$$

9.4 IMPEDANCE TENSOR

The relation between the electric and magnetic fields is described by the following equation, which can be represented in a matrix form as

$$\begin{bmatrix} E_x \\ E_y \end{bmatrix} = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{bmatrix} \begin{bmatrix} H_x \\ H_y \end{bmatrix} \quad (9.25)$$

The impedance Z is a complex tensor and is a function of the Earth's resistivity (ρ). The solution to the MT problem simplifies to reconstruction of resistivity from the frequency dependence of Z . The asymmetry of the material is given by on-diagonal components of Z in eq. 9.25. They become zero if the material is symmetric about a vertical plane passing through the observation site (Berdichevsky and Dimitriev, 1976). The off-diagonal components of Z give the information about vertical variations in resistivity structure. The spatial distribution of resistivity can be classified in three ways.

9.4.1 1-D EARTH

For 1-D Earth, we assume a horizontally layered Earth, where the conductivity σ only varies with depth along the z axis (Zhang et al., 1993). Maxwell's equations can be analytically solved by properly applying the boundary conditions. The solutions are electromagnetic waves (e.g. $\frac{\partial^2 E_x}{\partial z^2} = K^2 E_x$),

with the electromagnetic field always orthogonal to the magnetic field that travels perpendicular to the surface of the Earth in a constant oscillation direction. They attenuate with depending on their period and conductivity values. As a result, the MT transfer functions are independent of the orientation of the measured axes and are a function only of the frequency. The impedance tensor Z can be written as

$$Z = \begin{bmatrix} 0 & Z_{xy} \\ -Z_{yx} & 0 \end{bmatrix} \quad (9.26)$$

$$Z_{xy} = -Z_{yx} \neq 0.$$

For 1-D Earth, the apparent resistivity is given by

$$\rho_{xy} = \rho_{yx} = \frac{1}{\omega\mu} |Z|^2$$

$$\varphi_{xy} = \varphi_{yx} = \arg Z = \arg \sqrt{\frac{\omega\mu}{\sigma}} \sqrt{i} = \arg \sqrt{\frac{\omega\mu}{\sigma}} e^{i\frac{\pi}{4}} = \frac{\pi}{4} = 45^\circ$$

For homogeneous Earth, where an argument of the complex number $z = x + iy$, denoted by $\arg(z)$.

9.4.2 2-D EARTH

The figure in Zhang et al. (1993) illustrates a simplified two-dimensional (2-D) model with a vertical resistivity boundary striking in the x direction. The resistivity boundary may represent a geological fault or dyke. Electric currents on both sides of the interface must be conserved following the equation of continuity that the normal components of current density must be continuous on both sides. Ohm's law (eq. 9.4) can then be used to connect the current density to the electric field. For the y component,

$$j_{y_1} = \sigma_1 E_{y_1} = \sigma_2 E_{y_2} = j_{y_2} = \text{const.}$$

For two-dimensional Earth, resistivity varies with depth and in one horizontal direction. The direction in which the resistivity remains constant is called the geoelectric strike. The 2-D assumption requires that the strike length be significantly longer than the skin depth and all variations of the fields parallel to strike are zero ($\frac{\partial}{\partial x} = 0$). The coordinate system can be rotated in such way that $Z_{xx} = Z_{yy} = 0$ but $Z_{xy} \neq -Z_{yx}$. The impedance tensor becomes

$$Z = \begin{bmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{bmatrix} \quad (9.27)$$

The rotation of Z involves the decomposition of the electric and magnetic field components into E polarization (TE mode) and B polarization (TM mode), as explained by Berdichevsky and Dimitriev (1976). In the TE mode, the electric field is aligned with the electric strike, and in the TM mode, the magnetic field is aligned with the strike. If the x axis is in the strike direction, the off-diagonal components become

$$Z_{xy} = \frac{E'_x}{H'_y} = Z_{TE} \text{ (TE mode)}$$

$$Z_{yx} = \frac{E'_y}{H'_x} = Z_{TM} \text{ (TM mode)}$$

Often, Z_{TE} is written as $Z_{\parallel}$, that is, E_x is parallel to the strike, and Z_{TM} is written as $Z_{\perp}$, that is, E_y is perpendicular to the strike. The apparent resistivity for each mode is then

$$\rho_{xy} = \frac{1}{\omega\mu} |Z_{xy}|^2 = \frac{1}{\omega\mu} |Z_{TE}|^2$$

$$\rho_{yx} = \frac{1}{\omega\mu} |Z_{yx}|^2 = \frac{1}{\omega\mu} |Z_{TM}|^2$$

The phase angle in each mode for each component separately:

$$\varphi_{xy} = \arg Z_{xy} \text{ and } \varphi_{yx} = \arg Z_{yx}$$

9.4.3 3-D EARTH

In the 3-D Earth, the variations in resistivity occur in all three directions. It follows that the impedance tensor is

$$Z = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{bmatrix} \quad (9.28)$$

where $Z_{xx} \neq Z_{yy} \neq 0$ and $Z_{xy} \neq Z_{yx}$. There is no way to rotate the impedance tensor in such a way that the on-diagonal elements of Z become zero. Z is complex and can be separated into a real X and quadrature Y part:

$$\begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} = \begin{pmatrix} X_{xx} & X_{xy} \\ X_{yx} & X_{yy} \end{pmatrix} + i \begin{pmatrix} Y_{xx} & Y_{xy} \\ Y_{yx} & Y_{yy} \end{pmatrix}$$

TABLE 9.2

Entries of the Impedance Tensor Z vs Dimensionality and Isotropic Conductivity Distribution

	Dimensionality		
	1-D	2-D	3-D
Tensor components	$Z_{xx} = Z_{yy} = 0$ $Z_{xy} = -Z_{yx}$	$Z_{xx} = -Z_{yy}$ $Z_{xy} \neq -Z_{yx}$	$Z_{xx} \neq -Z_{yy} \neq Z_{xy} \neq -Z_{yx}$
Impedance tensor Z	$\begin{pmatrix} 0 & Z_n \\ -Z_n & 0 \end{pmatrix}$	$\begin{pmatrix} 0 & Z_{\parallel} \\ Z_{\perp} & 0 \end{pmatrix}$	$\begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix}$

Z has both magnitude and phase.

$$\rho_{a,ij}(\omega) = \frac{1}{\mu_0 \omega} |Z_{ij}(\omega)|^2$$

$$\varphi_{ij} = \tan^{-1} \left(\frac{\text{Im} Z_{ij}}{\text{Re} Z_{ij}} \right)$$

For a 3-D Earth, the resistivity varies in all three directions. All the impedance tensor elements in eq. 9.25 are different, and the components of the impedances are not equal and nonzero. Table 9.2 shows the entries of the impedance tensor Z vs dimensionality and isotropic conductivity distribution. The entries become increasingly independent for more complex dimensionality. $Z_{\parallel}$ and $Z_{\perp}$ denote the E-parallel and E-perpendicular impedance, respectively.

9.5 ROTATION OF IMPEDANCE TENSOR

Generally, the axes of the coordinate system in the field layout do not coincide with the strike direction. The impedance tensor can be rotated to any other coordinate system by an angle α with the rotation matrix R and its transpose R^T . The angle of rotation is in a clockwise direction from the x axis. The rotation operator is

$$R(\alpha) = \begin{bmatrix} \cos \alpha & \sin \alpha \\ -\sin \alpha & \cos \alpha \end{bmatrix}$$

The new impedance tensor related to new axes, x' and y' , and rotated by an angle α is given by

$$Z(\alpha) = R(\alpha) Z R(-\alpha)$$

The principal direction is defined as the direction when the off-diagonal components of $Z(\alpha)$ are maximized after rotation. Maximizing $|Z_{xy}(\alpha)|^2 + |Z_{yx}(\alpha)|^2$ gives

$$\alpha = \frac{1}{4} \arctan \frac{(Z_{xx} - Z_{yy})(\overline{Z_{xy} + Z_{yx}}) + (\overline{Z_{xx} + Z_{yy}})(Z_{xy} - Z_{yx})}{(Z_{xx} - Z_{yy})^2 - (Z_{xy} - Z_{yx})^2} \quad (9.29)$$

where α is the resulting strike angle, also known as the swift angle (Swift, 1967). The other way to determine principal directions is to minimize the on-diagonal components of $Z(\alpha)$, which yields the same result as in eq. 9.29.

9.6 MT STRIKE ANALYSIS

9.6.1 SKEW

Electrical strike reveals information about the directions in which the values of diagonal elements of impedance tensor come to a minimum. Often, it is the same direction as the geological fractures. The layout of the measurements rarely coincides with the strike direction. In order to perform two-dimensional analysis of MT data, the impedance tensor should be rotated by an angle α , as shown in Section 9.5. The angle it makes with geographical north is the swift angle (Swift, 1967). The rotation of Z by the angle α brings the impedance of the 3-D Earth close to the 2-D model. The strike angle can be determined from eq. 9.29 with respect to the initial coordinate system. However, this approach implies the 90° ambiguity in the strike angle, so there is no way to distinguish between α and $\alpha + 90^\circ$. The dimensionality of the structure is determined from the skew parameter:

$$S = \frac{|Z_{xx} + Z_{yy}|}{|Z_{xy} - Z_{yx}|}$$

The large value of S indicates the three-dimensionality of the subsurface structure. For 1-D Earth, the parameter S is zero, and for 2-D Earth, it should be close to zero after rotation of Z .

9.6.2 TIPPER

Another way to determine the strike direction was introduced by Jones and Vozoff (1978) and Everett and Hyndman (1967a, 1976b). The vertical component of magnetic field, H_z , may be expressed as a linear combination of horizontal magnetic field components H_x and H_y . It is given by

$$H_z = AH_x + BH_y$$

These complex coefficients can be visualized as operating on the horizontal magnetic field and tipping part of it into the vertical. The magnitude of the tipper is given by

$$T = \sqrt{A^2 + B^2}$$

The expressions for calculating the tipper coefficients are

$$A = \frac{\langle H_z H_x^* \rangle \langle H_y H_y^* \rangle - \langle H_z H_y^* \rangle \langle H_x H_x^* \rangle}{\langle H_x H_x^* \rangle \langle H_y H_y^* \rangle - \langle H_x H_y^* \rangle \langle H_y H_x^* \rangle}$$

$$B = \frac{\langle H_z H_y^* \rangle \langle H_x H_x^* \rangle - \langle H_z H_x^* \rangle \langle H_x H_y^* \rangle}{\langle H_y H_y^* \rangle \langle H_x H_x^* \rangle - \langle H_y H_x^* \rangle \langle H_x H_y^* \rangle}$$

where * denotes the complex conjugate. For a 2-D structure with strike in the x direction, $A = 0$. Tipper, thus, can be used to identify the presence of 2-D effects in the analyzed data. Information from the vertical magnetic field transfer functions is helpful in determining the structural strike direction (Radhakrishnamurthy and Likhite, 1970).

The angle which maximizes the coherency between the horizontal and vertical magnetic fields (Madden and Swift, 1969) is given by

$$\varphi = T^{-2} \left(\text{Re}^2 A + \text{Re}^2 B \right) \arctan(\text{Re} B + \text{Re} A) + (\text{Im}^2 A + \text{Im}^2 B) \arctan(\text{Im} B + \text{Im} A)$$

The phase difference between H_z and H_φ gives the bearing of some structures. They have also shown that the tipper skew γ is given by

$$\gamma = 2T^{-2} (\text{Re} A \text{Im} B - \text{Im} A \text{Re} B)$$

For 2-D structures, γ is zero. All three quantities T , φ , and γ are independent of axis rotation and provide some information about the subsurface structure. The reliability of the calculated H_z (H_z^c) is estimated from the coherence between it and the measured H_z (H_z^m). That is:

$$\text{Coh}(H_z^m H_z^c) = \frac{\left| A^* \langle H_z^m H_x^* \rangle + B^* \langle H_z^m H_y^* \rangle \right|}{\left\langle H_z^m H_z^{m*} \right\rangle^{\frac{1}{2}} \left[AA^* \langle H_x H_x^* \rangle + BB^* \langle H_y H_y^* \rangle + 2\text{Re} \left(AB^* \langle H_x H_y^* \rangle \right) \right]^{\frac{1}{2}}}$$

with * denoting the complex conjugate.

9.6.3 INDUCTION ARROWS

Induction arrows are vector representations of the complex (i.e. containing real and imaginary parts) ratios of the vertical to the horizontal magnetic field components. Since vertical magnetic fields are generated by lateral conductivity gradients, induction arrows can be used to infer the presence or absence of lateral variations in conductivity (Jupp and Vozoff, 1976). The vectors pointing towards the anomalous internal concentrations of current (Simpson and Bahr, 2005) are called the Parkinson convention, whereas the vectors pointing away from internal current concentrations are called the wise convention (Parkinson, 1959). The vectors are also sometimes called tipper vectors as they transform horizontal magnetic fields into the vertical plane according to the relationship:

$$H_z(\omega) = \begin{pmatrix} T_x(\omega) & T_y(\omega) \end{pmatrix} \begin{pmatrix} \frac{B_x}{\mu_0} \\ \frac{B_y}{\mu_0} \end{pmatrix}$$

In a 2-D Earth, induction arrows are associated only with the E polarization. Thus, insulator–conductor boundaries extending through a 2-D Earth give rise to induction arrows that orient perpendicular to them and have magnitudes that are proportional to the intensities of anomalous current concentrations (Wiese, 1962), which are, in turn, determined by the magnitude of the conductivity gradient or discontinuity. However, an absence of induction arrows at a single site does not necessarily confirm an absence of laterally displaced conductivity boundaries (Jupp and Vozoff, 1976).

9.7 THE EFFECTS OF STATIC SHIFT

The MT method, which measures an electric field on the surface, suffers from the telluric or static shift problems. The effect of static shift occurs due to the presence of small-scale near-surface inhomogeneities, and it appears as a vertical shifting of the apparent resistivity curve by a frequency-independent factor without any change in the phase curve (Sasaki, 2004). Jiracek (1990) proposed that the shift is caused by resistivity inhomogeneities in the vicinity of the surveyed area and especially close to the electrical dipoles. The reasons are the voltage distortion of the electric field by highly conductive minerals, topographic distortion, and current channelling. The voltage distortion occurs as a result of local resistivity anomaly near the surface, the topographic distortion takes place under the hills and valleys affecting the current density, and current channelling occurs when the current is deflected by resistivity anomaly. Jiracek (1990) gives an overview of the static shift problem and proposes the methods for its correction.

9.8 3-D GALVANIC DISTORTION AND DECOMPOSITION OF MT IMPEDANCE TENSOR

The electromagnetic signal becomes weak when the electromagnetic skin depth becomes greater than the dimensions of the anomaly, but it continues to have a noninductive response, termed as *galvanic* (Jones and Price, 1970).

Electromagnetic data containing galvanic effects can often be described by the superimposition or decomposition model, in which the data are decomposed into a noninductive response owing to multidimensional heterogeneities with dimensions significantly less than the inductive scale length of the data (often described as local) and a response owing to an underlying 1-D or 2-D structure (often described as regional). In such cases, determining the electromagnetic strike involves decomposing the measured impedance tensor into matrices representing the inductive and noninductive parts. The inductive part is contained in a tensor composed of components that have both magnitude and phase (i.e. its components are complex), whereas the noninductive part exhibits DC behaviour only and is described by a distortion tensor, the components of which must be real and, therefore, frequency-independent (Jones and Price, 1970).

In summary, the purpose of our decomposition is to separate local and regional parameters as much as possible under the assumption that the regional structure is, at most, 2-D and the local structure causes only galvanic scattering of the electric fields, and to do so in the form of a product factorization.

The decomposition technique, which is used to determine the electromagnetic strike using the decomposition hypothesis, is an inverse technique proposed by Groom and Bailey (Groom and Bailey, 1989). In the Groom–Bailey decomposition technique, the separation of the ‘localized’ effects of 3-D current channelling from the ‘regional’ 2-D inductive behaviour is achieved by factorizing the impedance tensor problem in terms of a rotation matrix, β , and a distortion tensor, C , which is itself the product of three tensor suboperators (twist T , shear S , local anisotropy A , and a scalar g) (Simpson and Bahr, 2005):

$$Z = \beta C Z \beta^T \text{ where } C = gTSA$$

$$\text{Where } \beta = \begin{pmatrix} \cos\alpha & \sin\alpha \\ -\sin\alpha & \cos\alpha \end{pmatrix} \text{ and } \beta^T = \begin{pmatrix} \cos\alpha & -\sin\alpha \\ \sin\alpha & \cos\alpha \end{pmatrix}$$

9.9 ROTATION-INVARIANT MAGNETOTELLURIC IMPEDANCE TENSORS

A clear step forward is taken by more general approaches which require no 2-D assumptions of the underlying structure. A separate class of impedance tensors known as rotation-invariant tensors can also be used for interpretation of the MT data. Rotation-invariant MT parameters are those whose

amplitudes remain invariant as the MT tensors are rotated mathematically. In a layered Earth, the impedance in all directions can be calculated (from the bivariate regression) simply by measuring the electric field variation in the perpendicular direction. We could test the hypothesis of a layered Earth by applying a mathematical rotation to the impedance tensor estimated from data measured in a fixed coordinate frame.

Theoretically, we can simulate a field site setup with sensors oriented in any direction, α , via a mathematical rotation involving matrix multiplication of the measured fields (or impedance tensor) with a rotation matrix, β (Jupp and Vozoff, 1976):

$$\mathbf{Z}' = \beta \mathbf{Z} \beta^T$$

$$\text{where } \beta = \begin{pmatrix} \cos\alpha & \sin\alpha \\ -\sin\alpha & \cos\alpha \end{pmatrix}$$

and superscript T denotes the transpose of β , $\beta^T = \begin{pmatrix} \cos\alpha & -\sin\alpha \\ \sin\alpha & \cos\alpha \end{pmatrix}$.

There is an infinite number of these rotation-invariant tensors (Szarka and Menville, 1997; Eggers, 1982; Yee and Paulson, 1987). Out of these tensors, $(\rho_{\text{determinant}}, \varphi_{\text{determinant}})$ and $(\rho_{\text{average}}, \varphi_{\text{average}})$, as proposed by Berdichevsky (1968), Berdichevsky and Dimitriev (1976), and $(\rho_{\text{central}}, \varphi_{\text{central}})$, as proposed by Lilley (1976, 1993), can be used for interpretation of MT data because the information content in these parameters is more and their apparent resistivity and phase values are comparable with those obtained from the TE and TM mode apparent resistivities and phases. A detailed study of the properties of the rotation-invariant parameters (Roy et al., 1998) reveals that three pairs of rotation-invariant parameters can be used for interpretation of subsurface structures because their values are quite close to each other in most of the frequencies and they remain within the upper and lower bounds of $(\rho_{\text{axy}}, \varphi_{\text{axy}})$ and $(\rho_{\text{ayx}}, \varphi_{\text{ayx}})$ near a major contact of a 2-D or nearly 2-D geological structure. All the real and imaginary components of the four impedance tensors $Z_{\text{xy}}, Z_{\text{yx}}, Z_{\text{xx}},$ and Z_{yy} are present in $(\rho_{\text{determinant}}, \varphi_{\text{determinant}})$ and $(\rho_{\text{central}}, \varphi_{\text{central}})$. The expressions of these three pairs of rotation-invariant tensors are given as:

$$\rho_{\text{determinant}} (\rho_D) = \frac{1}{\omega \mu_0} [Z_{\text{xx}} Z_{\text{yy}} - Z_{\text{xy}} Z_{\text{yx}}]$$

$$\varphi_{\text{determinant}} (\varphi_D) = \text{phase of } \frac{Z_{\text{xx}} Z_{\text{yy}} - Z_{\text{xy}} Z_{\text{yx}}}{2}$$

$$\rho_{\text{average}} (\rho_B) = \frac{1}{\omega \mu_0} [Z_B]^2 \text{ where } Z_B = \frac{Z_{\text{xy}} - Z_{\text{yx}}}{2}$$

$$\varphi_{\text{average}} (\varphi_B) = \text{phase of } \frac{Z_{\text{xy}} - Z_{\text{yx}}}{2}$$

$$\rho_{\text{central}} (\rho_C) = \frac{1}{\omega \mu_0} [Z_C]^2 \text{ where } Z_C = [X^2 + Y^2]^{1/2}$$

$$\varphi_{\text{central}} (\varphi_C) = \tan^{-1} \frac{Y}{X}$$

$$X = \frac{1}{2} [(Z_{\text{xxr}} + Z_{\text{yyr}})^2 + (Z_{\text{xyr}} - Z_{\text{yxr}})^2]^{1/2} \text{ and } Y = \frac{1}{2} [(Z_{\text{xxi}} + Z_{\text{yyi}})^2 + (Z_{\text{xyi}} - Z_{\text{yxi}})^2]^{1/2}$$

where r and i stand for the real and imaginary components, respectively, of the impedance tensor element Z (Lilley, 1993). Here, ω is the angular frequency of the MT signals and μ_0 is the free space magnetic permeability.

Once the impedances are computed, one can calculate a quantity called the invariant, which is independent from the orientation of the MT array relative to the underlying structures. The invariant is equal to

$$Z_{xx} Z_{yy} - Z_{xy} Z_{yx}$$

Rather than plotting the invariant versus the period, one prefers to compute an apparent resistivity.

$$\varrho_{\text{invariant}} = \frac{1}{\omega \mu_0} (Z_{xx} Z_{yy} - Z_{xy} Z_{yx})$$

Depending upon the algorithm used to compute the corresponding impedances, this quantity will be under- or overestimated or correct.

9.10 MAGNETOTELLURIC PHASE TENSORS

Caldwell et al. (2004) introduced another elegant method that did not require presumptions about the underlying dimensionality. That means, the MT phase tensor is applicable where both the regional conductivity distribution and the conductivity heterogeneity close to the surface are 3-D. The tangent of the MT phase was defined in eq. (9.22) as the ratio of the imaginary and real parts of the impedance tensor Z . This expression allows a generalization to the entire tensor, specifically its real and imaginary parts X and Y , respectively.

$$\varphi = X^{-1}Y$$

Here, φ defines the real phase tensor and X^{-1} is the inverse of the real part of the impedance tensor Z . In order to follow the considerations about the regional and galvanic distortion of the electromagnetic fields, the phase tensor φ can also be expressed in terms of the regional field only.

$$\varphi = X^{-1}Y = (CX_R)^{-1}(CY_R) = X_R^{-1}C^{-1}CY_R = X_R^{-1}Y_R = \varphi_R$$

As expected, the phase behaves independent of the distortion C near the surface, which affected the observation of the impedance tensor.

Figure 9.1 illustrates the phase tensors with its three invariants. The three coordinate invariants are the maximum phase tensor value $\varphi_{\max}$, the minimum phase tensor value $\varphi_{\min}$, and the skew angle β . The skew angle defines the asymmetry of the tensor and deviates from the symmetry axis (dashed line in Figure 9.1). The angle α expresses the tensor's dependence on the coordinate system. However, α is not coordinate-invariant.

$$\alpha = \frac{1}{2} \tan^{-1} \left(\frac{\varphi_{xy} - \varphi_{yx}}{\varphi_{xx} + \varphi_{yy}} \right)$$

The phase tensor is fully defined with the angle α and the three coordinate invariants.

$$\varphi = R^T (\alpha - \beta) \begin{pmatrix} \varphi_{\max} & 0 \\ 0 & \varphi_{\min} \end{pmatrix} R (\alpha + \beta)$$

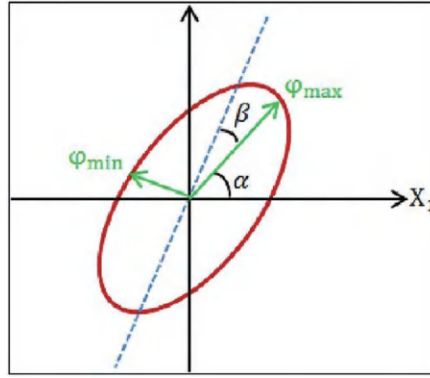


FIGURE 9.1 The MT phase tensor φ graphically illustrated as an ellipse. The major and minor axes of the ellipse are proportional to the principal values of φ (namely, $\varphi_{\max}$ and $\varphi_{\min}$), which define the tensor's dependence on the coordinate system (x, y) . If the skew angle $\beta = 0$, the principal values are equal to the eigenvalues of the tensor and $\varphi_{\max}$ is aligned with the symmetry axis (dashed line). The direction of the major axis $\varphi_{\max}$ is defined by $(\alpha - \beta)$.

Source: Caldwell et al. (2004).

where $R(\alpha + \beta)$ is a rotation matrix, with its transpose or inverse $R^T(\theta) = R(-\theta)$:

$$R(\alpha + \beta) = \begin{pmatrix} \cos(\alpha + \beta) & \sin(\alpha + \beta) \\ -\sin(\alpha + \beta) & \cos(\alpha + \beta) \end{pmatrix}$$

The orientation of the major axis of the tensor is $\alpha - \beta$. If $\beta = 0$, the tensor is symmetric.

9.11 INVERSION ROUTINES

9.11.1 OVERVIEW OF INVERSION METHODS

The inverse models are discretized into M constant resistivity blocks $m = [m_1, m_2, \dots, m_M]$, and there are N observed data $d = [d_1, d_2, \dots, d_N]$ with uncertainties $e = [e_1, e_2, \dots, e_N]$. The data misfit functional φ_d is defined as

$$\varphi_d = (d - F[m])^T C_d^{-1} (d - F[m]) \quad (9.30)$$

where $F[m]$ denotes the model response and C_d the data covariance matrix, which is diagonal and contains the data uncertainties e . In principle, the inverse problem formulated in eq. (9.30) will converge. However, the inverse problem is non-unique due to data errors, especially under sampled data, limited-frequency bands, and assumptions about dimensionality (Parker, 1980). Therefore, a model structure functional φ_m is introduced:

$$\varphi_m = (m - m_0)^T C_m^{-1} (m - m_0) \quad (9.31)$$

with C_m the model covariance matrix, which characterizes the magnitude and smoothness of resistivity variation with respect to the prior model m_0 . The two functionals are combined to yield the unconstrained functional φ_λ with X_* the desired level of misfit:

$$\varphi_\lambda(m) = \varphi_m + \lambda^{-1} (\varphi_d - X_*^2)^{m, \lambda} \rightarrow \min \quad (9.32)$$

The Lagrange multiplier λ controls whether more weight is given to minimizing the norm of data misfit or the norm of the model (Tikhonov and Arsenin, 1977). For larger values of λ , the data misfit becomes less important and more weight is given toward producing a smoother model. In order to minimize eq. 9.32, the stationary points have to be found. Instead of using eq. 9.32, the penalty functional $W_\lambda(m)$ is differentiated with respect to m :

$$W_\lambda(m) = \varphi_m + \lambda^{-1} \varphi_d \xrightarrow{m, \lambda} \min \quad (9.33)$$

For a constant value of λ , the stationary points of $\varphi_\lambda(m)$ and $W_\lambda(m)$ are the same. Therefore, minimizing $W_\lambda(m)$ with varying λ gives the stationary points of $\varphi_\lambda(m)$, and a value for λ can be found such that the data misfit φ_d is equal to X_*^2 .

9.11.2 THE MODEL SPACE OCCAM AND NON-LINEAR CONJUGATE GRADIENT (NLCG) METHODS

The MT inverse problem is non-linear (that means $F[m]$), and therefore, iterative solutions to solving eq. 9.33 are required. The model response $F(m)$ is approximated by a Taylor series expansion:

$$F(m_{k+1}) = F(m_k + \Delta m) = F(m_k) + J_k(m_{k+1} - m_k) \quad (9.34)$$

The subscript k is the iteration number, and $J = \left(\frac{\partial F}{\partial m} \right)_k$ is the $N \times M$ Jacobian or sensitivity matrix.

In order to find the stationary points, eq. 9.34 is substituted into eq. 9.33, which yields iterative solutions:

$$m_{k+1}(\lambda) = (\lambda C_m^{-1} + T_k^m)^{-1} J_k^T C_d^{-1} X_k + m_0 \quad (9.35)$$

With $X_k = d - F(m_k) + J_k(m_k - m_0)$, and the $M \times M$ model space cross-product matrix $T_m^k = J_k^T C_d^{-1} J_k$ (Siripunvaraporn and Egbert, 2000; Siripunvaraporn et al., 2005a).

The Occam inversion runs through two different stages before it reaches a final solution. The Lagrange multiplier λ is used both as a step length control and a smoothing parameter. A series of values for λ are used to minimize the misfit of each iteration (eq. 9.35) in phase 1. The aim is to reduce the misfit to the desired level X_*^2 . However, in the early iterations, the misfit level X_*^2 will not be reached for a series of values for λ . Each stage of iteration involves the Occam code using the minimum misfit from the best λ as a basis for the next iteration. Once the desired misfit X_*^2 is reached, phase 2 commences by keeping the misfit at the desired level and varying λ to search for the smallest norm, that is, the smoothest model. The downside of the Occam inversion is that the sensitivity matrix J must be calculated for each iteration (eq. 9.35). This process is both time-consuming and requires much computer memory to store J .

The NLCG method used by Rodi and Mackie (2001) offers an improvement in speed due to pre-conditioning and the computation of only the gradient $g = \frac{\partial \varphi}{\partial m}$ rather than the sensitivities J . Here, the Lagrange multiplier λ is kept fixed, and a suitable value for λ must be found for the inversion to find a model that satisfies both the data misfit and the model structure functional. In contrast to the Occam approach, the NLCG method requires the solution of a univariate minimization problem at each stage of iteration. This is also referred to as line search, which has been optimized by Rodi and Mackie (2001) to take only as long as the computation of three solutions of the forward problem. The NLCG modelling approach is therefore quicker than the Occam inversion, which requires the computation of the entire sensitivity matrix at each stage of iteration.

9.11.3 THE DATA SPACE METHOD

Following the argument of Parker (1980), the iterative solution eq. 9.35 can be reformulated as a linear combination of rows of the smoothed sensitivity matrix $C_m J^T$:

$$m_{k+1} - m_0 = C_m J_k^T \beta_{k+1} \quad (9.36)$$

Here, the unknown expansion coefficient vector of the ‘basis functions’ $[C_m J_k^T]_j$ with $j = 1, \dots, N$ is called β_{k+1} . Eq. 9.36 can also be substituted into eq. 9.33 and solve for stationary points, which will give another set of iterative solutions:

$$\beta_{k+1} = (\lambda C_d + T_k^n)^{-1} X_k \quad (9.37)$$

where $T_k^n = J_k C_m J_k^T$ is the $N \times N$ data space cross-product matrix. Similar to the Occam approach, λ is used to minimize the misfits among the solutions of eq. 9.37 to a desired misfit level, followed by a search for the smallest norm. The solutions from both the model space and the data space approaches are, in theory, identical (Siripunvaraporn et al., 2005a); however, depending on the model and data size M and N , respectively, the number of equations to be solved can be significantly different. The data space dimensions $N \times N$ are usually much smaller in practice (especially in 3-D inverse problems), reducing the computational costs. Siripunvaraporn et al. (2005a) note that the model covariance C_m is used in the data space approach, but its inverse C_m^{-1} is implemented in the model space approach. The advantage of using the model covariance itself is that a priori information such as the ocean can be readily included. It should be noted that the model covariance C_m itself is never computed, but only the product with the sensitivity matrix $C_m J^T$ is required.

9.12 VARIANT OF MAGNETOTELLURICS

Audio magnetotellurics (AMT) is a higher-frequency magnetotelluric technique for shallower investigations. While AMT has less depth penetration than MT, AMT measurements often take only about 1 hour to perform (but deep AMT measurements during low-signal-strength periods may take up to 24 hours) and use smaller and lighter magnetic sensors. Transient AMT is an AMT variant that records only temporarily during periods of more intense natural signal (transient impulses), improving signal-to-noise ratio at the expense of strong linear polarization.

CSEM controlled source electromagnetic is a deepwater offshore variant of controlled source audio magnetotellurics; CSEM is the name used in the offshore oil and gas industry. Onshore CSEM/CSAMT may be effective where electromagnetic cultural noise (e.g. power lines, electric fences) presents interference problems for natural-source geophysical methods. An extensive grounded wire (2 km or more) has currents at a range of frequencies 0.1 Hz to 100 kHz passed through it. Lateral resolution is determined by station spacing. Station spacing is typically between 10 and 200 m. The received signal strength is proportional to the size of the station spacing. For example, if the station size is cut in half, the signal strength is cut in half. This is important to remember when designing a survey, because the limiting factor on depth of exploration is usually the signal strength. The CSAMT method has proven useful for mapping the Earth’s crust in the 20 to 1,000 m depth range. Vertical resolution is 5–20% of the depth. The electric field parallel to the source and the magnetic field which is at right angles are measured. The resistivity is then calculated, and the lower the resistivity, the more likely there is a conductive target (graphite, nickel ore, or iron ore). CSAMT is also known in the oil and gas industry as onshore controlled source electromagnetics (onshore CSEM).

An offshore variant of MT, the marine magnetotelluric (MMT) method, uses instruments and sensors in pressure housings deployed by ship into shallow coastal areas where water is less than 300 m deep. A derivative of MMT is offshore single-channel measurement of the vertical magnetic

field only (the Hz, or ‘tipper’), which eliminates the need for telluric measurements and horizontal magnetic measurements. While the theory is sound and many case histories exist, no commercial systems are available. Furthermore, any such system would require a solution providing for the precise orientation and stabilization of the magnetic sensor.

9.13 INSTRUMENTS USED IN MT

MT instrumentation design and construction are a specialized international activity, with only a small number of companies and scientific organizations having the necessary expertise and technology. Three companies supply most of the commercial-use world market:

- One in the United States (KMS Technologies¹ and Zonge International Inc.)²
- Two in Canada (Advanced Geophysical Operations and Services Inc. [AGCOS]³ and Phoenix Geophysics Ltd.)⁴
- One in Ukraine (LEMI LLC, a joint venture with KMS Technologies)⁵
- One in Germany (Metronix Messgeraete und Elektronik GmbH)⁶

High-power CSEM array systems above 100 KVA for hydrocarbon and geothermal exploration are manufactured by KMS Technologies, USA. Government agencies and smaller companies producing MT instrumentation for internal use include Vega Geophysics Ltd.⁷ in Russia and the Russian Academy of Sciences (SPbF IZMIRAN) and the National Space Research Institute of Ukraine. Geometrics Inc. of San Jose, California, USA, produces a high-frequency, hybrid-source AMT system combining both natural fields and a high-frequency (1–70 kHz) transmitter, as well as a multichannel distributed network CSAMT instrument.

Let us discuss with a necessary example of MT instrument (24-Bit Geophysical Measuring System GMS-06 from Metronix) for easy comprehension. Only a few components are required to set up an MT station (Metronix, Germany):

- A five-channel ADU-06
- Three magnetic field sensors MFS-06 with cables (typically induction coil sensors)
- Four e-field probes with cables (telluric/electric sensors)
- One network cable (temporary)
- One laptop for station setup and data control (not necessarily required during measurement)
- One 12 V battery

The main components of ADU-06 are the network board (PC/104), CPU board (PC/104), 2–5 A/D boards (PC/104), GPS module and GPS antenna, clock control board, power supply board, connector board, flash disk, and optional hard disk.

Technical data of ADU-06 include frequency range: DC to 20 KHz; number of channels: 2, 3, or 5 channels; A/D conversion: 24-bit; system computer: 32-bit controller; storage media: flash disk 256 MB (standard) up to 2 GB (option); synchronization: GPS clock ± 130 ns or better; interfaces: network, magnetometers, e-lines, battery input, GPS antenna; weight and dimensions: approximately 6 kg, 23 cm $\times$ 20 cm $\times$ 18 cm; operating temperature: -40°C to 70°C (extended) or -15°C to 60°C (standard).

Details about MT instrumentation can be found from different sites of the relevant agencies/companies, as follows.

9.14 MT FIELD LAYOUT

One uses a classical five-component array (Metronix Germany; Naskar and Das, 2015) and records the variations of the electrical field in two horizontal directions at right angles (E_x , E_y). One then measures the potential difference which exists between two electrodes of a dipole and

records the variations of the magnetic field in three directions at right angles, two of them, H_x (or B_x) and H_y (or B_y), being parallel to the electrical dipoles ($H_x \parallel E_x$ and $H_y \parallel E_y$), the third one being vertical.

Length of potential dipole is usually in the order of 200 m; the longer the spread length, the better the coverage. However, practical considerations limit their length, which seldom exceeds 300 m (presence of roads, topographic effects, presence of possible noise source in their vicinity, such as electrified fences, buried pipes). Electrodes are of impolarizable type, filled with Pb-PbCl_2 (lead–lead chloride). In order to eliminate any noise generated by the movement of the dipole cable by the wind, pods of Earth are placed every 1 m apart on the cable (a conductive cable moved in the Earth's magnetic field induces a current). The equipment measures the difference of potential existing between two electrodes (in mV). Magnetometers are induction type. They comprise 600,000 turns of copper wire wounded around a core of high magnetic permeability (μ -metal). They have a very flat response over a broad frequency range due to a special feedback system and are equipped with a preamplifier. The magnetometer is a cylinder 1.8 m long and 15–20 cm wide. The horizontal magnetometers are carefully oriented and levelled within a fraction of degree. They are buried in trenches 50–80 cm deep. The vertical magnetometer is placed vertically in an anger hole and levelled with the same care as the horizontal magnetometers. All these precautions are taken to eliminate any source of noise arising from thermal variations and vibrations (wind, for instance). Negligible electronic noise is in the 10–100 Hz range; the noise is only due to the thermic noise of the coil itself. This is a frequency range where the signal is frequently weak. Above 20 seconds, the natural signal is usually strong enough for the instrumental noise to be negligible. To conclude, for horizontal components and periods of practical interest for oil exploration, MT recording equipment is associated with:

- Low-pass filters 0.2 Hz to 0.005 Hz
- High-pass filters 100 Hz
- Notch filters 50 Hz and 150 Hz or 60 Hz and 180 Hz
- Amplification gains 1-3-10-30

And selecting the frequency range (HF = 0.2 to 128 Hz, MF = 0.2 to 16 Hz, LF = 0 to 1 Hz or 0.1 to 0.0001 Hz).

A complete five-component set of MT equipment can be backpack-carried by a small field team (two to four persons) or carried by a light helicopter (such as the MD helicopters MD 500), allowing deployment in remote and rugged areas. Most MT equipment is capable of reliable operation over a wide range of environmental conditions, with ratings of typically -25°C to $+55^\circ\text{C}$, from dry desert to high humidity (condensing) and temporary full immersion.

9.14.1 MULTICHANNEL CONFIGURATION

In the multichannel configuration, multiple ADUs can be operated either in a network or as stand-alone stations (synchronization by GPS signal) (Metronix Germany).

9.14.2 REMOTE REFERENCE MAGNETOTELLURIC (RRMT)

Remote reference is an MT technique used to reduce the bias caused by noise in electric field E and magnetic field H at a local site and by acquiring simultaneous data at more than one MT station. Remote reference (RR) only works when noise at a local and remote site is uncorrelated. This greatly improves data quality and may allow acquisition in areas where the natural MT signal is difficult to detect because of man-made EM interference.

9.15 PROCESSING AND INTERPRETATION OF MT DATA

Horizontal orthogonal components of time-varying electric field and all the three components of the magnetic field are recorded as time series. These time series are fast Fourier transformed (FFT) to estimate the impedance (Z), which relates the observed electric field (E) to the magnetic field (H). The information about the conductivity structure of the Earth is encoded in the Z as a function of frequency. The correct estimation of Z is the crux of the problem in MT method. The presence of noise in the magnetic field will result in biased estimates of impedance $E = ZH$; Z is impedance. Impedance is simply related to an apparent resistivity. The use of a tensor impedance relationship is necessitated by the distortion imposed on the electromagnetic field by lateral variations in subsurface resistivity. The phase of the measured electric field over the magnetic field provides useful information of the conductivity response.

The data can be interpreted in all the 1-D, 2-D, and 3-D domains using the different inversion approaches. Normally, singular value decomposition (SVD), ridge regression (RR), weighted ridge regression (WRR), conjugate gradient minimization (CGM), and stochastic inversion (SI) are some of the approaches used for linearized inversion, where sensitivity matrix is used. Monte Carlo method (MCM), simulated annealing (SA), very fast simulated annealing (VFSA), and genetic algorithm (GA) are the global optimization tools used quite often for interpretation of MT data. Artificial neural network method has (ANNM) also been used in quite a few cases. For 2-D subsurface, rotation-invariant tensors cannot be used theoretically because the Helmholtz electromagnetic wave equations which generate TE and TM mode EM equations after optimum rotation are not applicable to rotation-invariant tensors. For 2-D and 3-D forward problems, the finite difference (FD), finite elements (FE), and integral equation (IE) methods are used widely.

9.16 APPLICATIONS OF MT METHOD AND CASE STUDY

Several academic institutions and research laboratories are equipped with MT data acquisition, processing, and multidimensional modelling and inversion facilities. The Indian groups involved in MT field investigations and theoretical development works, along with their contributions, software, are listed in Table 9.3.

From north to south, the various groups actively pursuing MT studies include GSI, WIHG, IITR, KUK, ISM(IIT), ISR, GERMI, IITB, IIGM, AU, OU, CSIR-NGRI. These groups, namely, the Geological Survey of India (GSI), Indian Institute of Technology Roorkee (IITR), Kurukshetra University Kurukshetra (KUK), Council of Scientific and Industrial Research National Geophysical Research Institute (CSIR-NGRI), Indian Institute of Geomagnetism Mumbai (IIGM), Institute of Seismological Research (ISR, Gandhinagar), Andhra University (AU), Gujarat Energy Research and Management Institute (GERMI), Osmania University (OU), Indian School of Mines (ISM:IIT, Dhanbad), Indian Institute of Technology Bombay (IITB), Union Carbide India Limited (UCIL), and Oil and Natural Gas Corporation (ONGC) have worked in different parts of India. Locations of MT sites distributed along different profiles are shown on the simplified tectonic map of India in the figure in Israil et al. (2020; modified after Kusham et al., 2018).

The numeral in the figure in Israil et al. (2020; modified after Kusham et al., 2018) indicates both the relevant reference and the profile/region. The tectonic regions covered by these MT investigations are shown in the figure in Israil et al. (2020; modified after Kusham et al., 2018). The field investigations aimed to delineate deep crustal structure, the basement depth geometry, fresh-water-bearing aquifer and contaminated zone, thickness of the Gondwana and Siwalik sediments, rifting of the Narmada–Son lineament (NSL) zone, resistivity structure in the Bhuj earthquake region, electrical resistivity characterization of fault zones, mapping of geothermal source zone,

and lithosphere architecture. The software AP3DMT-MATLAB code for 3-D inversion of MT data was developed and tested. This code is made versatile and user-friendly by an efficient use of MATLAB's inbuilt functions.

TABLE 9.3
List of MT Survey Groups in India

Study Area	Working Group	Major Results
Himalaya	GSI, IITR, KUK, CSIR-NGRI	Existence of transverse tectonic features along with thickness of the sedimentary layers has been discussed.
Indo-Gangetic Plain (IGP)	GSI, CSIR-NGRI	Thickness of sediments, freshwater-bearing aquifer zone, and basement topographic variation are estimated.
Gujarat	ISR, CSIR-NGRI, IIGM, KUK, AU, GERMI	MT investigations in and around hot water spring. High-conductive thermal influx of mantle exchange of heat to lower crust and sedimentary thicknesses were delineated.
Maharashtra	GSI, CSIR-NGRI, OU	Geothermal region, processing of culturally noise-contaminated data of Koyana–Warna seismic zone, identification of the variation of the thickness of Deccan volcanics, nature of the granitic basement, and lithospheric electrical structure in the Western Ghat were presented.
Eastern India	ISM (IIT), IITB, UCIL	Resistivity structure was used to confirm the southward extension of Eastern Ghat Mobile Belt (EGMB). Low-resistivity zones were correlated with the presence of gold, silver, uranium, and copper deposit in the study area.
Central India	GSI, CSIR-NGRI	Rifting of the Narmada–Son lineament (NSL) zone and extension of igneous body of large igneous province of Deccan volcanics in the NSL region were explained on the basis of resistivity model.
Dharwar	CSIR-NGRI	Evidence for mantle suture and plume–continental interaction, hotspots and the metasomatized fluids inferred by low resistivity near Billigiri Rangan massif, and resistivity structure of kimberlite melts zone in Dharwar craton are discussed.
South India	IIGM, CSIR-NGRI, ONGC	Dipping resistive crustal layer in the NE parts of Cuddapah basin inferred as E–W compression along the eastern margin during the Neoproterozoic–Neoproterozoic tectonic convergence between India and East Antarctica were discussed.
Software	IITR	MATLAB-based computer code, AP3DMT, for inversion of 3-D magnetotelluric (MT) data was developed and tested over a variety of models.

Source: Modified after Israil et al. (2020).

9.16.1 COMMERCIAL APPLICATIONS

9.16.1.1 Hydrocarbon Exploration

For hydrocarbon exploration, MT is mainly used as a complement to the primary technique of reflection seismological exploration. While seismic imaging is able to image subsurface structure, it cannot detect the changes in resistivity associated with hydrocarbons and hydrocarbon-bearing formations. MT does detect resistivity variations in subsurface structures, which can differentiate between structures bearing hydrocarbons and those that do not.

At a basic level of interpretation, resistivity is correlated with different rock types. High-velocity layers are typically highly resistive, whereas sediments—porous and permeable—are typically much less resistive. In contrary to that fact that high-velocity layers are an acoustic barrier and make seismic ineffective, their electrical resistivity means the magnetic signal passes through almost unimpeded. This allows MT to see deep beneath these acoustic barrier layers, complementing the seismic data and assisting interpretation. 3-D MT survey results in Uzbekistan (32×32 grid of soundings) have guided further seismic mapping of a large known gas-bearing formation with complex subsurface geology.⁸ China National Petroleum Corporation (CNPC) and Nord-West Ltd. (www.nw-geophysics.com) use onshore MT more than any other oil company in the world, conducting thousands of MT soundings for hydrocarbon exploration and mapping throughout the globe.⁹

9.16.1.1.1 Saurashtra Region, Gujarat, India

In the north-western region of the Saurashtra peninsula, CSIR-NGRI initially started MT studies during 1987–1990 along two NS and EW trending profiles (Sarma et al., 1992). These studies have given information about subtrappean sedimentary formation in the northern part of the Saurashtra region. Later, Oil and Natural Gas Corporation Limited (ONGC) drilled a borehole near Lodhika. The borehole results are well coinciding with the MT results, showing the subtrappean sedimentary layer beneath the Saurashtra region. Consequently, CSIR-NGRI received a major project for integral geophysical studies in the Saurashtra region from the Oil Industry Development Board (OIDB) and ONGC. In this sponsored project, CSIR-NGRI acquired 600 MT sites in a grid pattern in the NW part of Saurashtra near the Lodhika borehole for exploration of Mesozoic sediments (Sarma et al., 1998). The modelling of the entire dataset successfully delineated the major Mesozoic sedimentary formation beneath the NW part of the Saurashtra peninsula (Patro, 2017). 3-D inversion of MT data along an NS profile (Patro et al., 2021, profile no. 6 for location) delineated high-resistive ($>4,000 \Omega\text{-m}$) granitic basement extending up to 20–25 km (Patro et al., 2021). Within the crustal column, various faults, such as the Rajula fault (RF) and the Saverkindala fault (SKF), have been delineated, and they have the same geological trend as that of the Narmada–Son lineament (NSL) (Ajithabh and Patro, 2021). Ajithabh and Patro (2021) inferred the extension of NSL to Saurashtra by analyzing these datasets together with geological and seismological studies. This success has paved the way for the initiation of several MT exploration programs in the country for hydrocarbon exploration by integrating MT results with gravity and seismic reflection (Sarma et al., 1992, 1998).

9.16.1.2 Mineral Exploration

MT is used for various base metals (nickel and precious metals exploration, as well as for kimberlite) mapping. INCO's 1991 proof-of-concept study in Sudbury, Ontario, Canada, sensed a 1,750 m deep nickel deposit. Falconbridge followed with a feasibility study in 1996 that accurately located two Ni-Cu mineralized zones at about 800 m and 1,350 m depth. Since then, both major and junior mining companies are increasingly using MT and AMT for both brownfield and greenfield explorations. Significant MT mapping work has been done on areas of the Canadian Shield.¹⁰ Diamond exploration, by detecting kimberlites, is a proven application.¹¹

9.16.1.2.1 Cuddapah Basin, Southern India

The IIGM, CSIR-NGRI groups conducted an MT survey in the Cuddapah basin (Southern India) (Konda et al., 2015). The geoelectric structures of Palnad Sub and Nallamalai Fold Belt, Cuddapah Basins, were estimated along an NW–SE profile extending between Potuluru and Anjanpalli in the NE part of Cuddapah basin along the profile (Israil et al., 2020, profile ‘23’). The resistivity depth model beneath the profile shows a thick low-resistive (40–300 Ωm) zone, representing sediments of the Palnad sub-basin (PSB) with thickness varying between 350 and 2,000 m. The low resistivity is interpreted as the presence of base metals (Cu, Pb, and Zn) within the Banaganapalli quartzite and Narji limestone. A moderately low resistive feature (450–800 Ωm), dipping south-east in the centre of the profile, represents Palnad sub-basin sediments entrapped along the fault zone, which coincides with the Nallamalai Fold Belt (NFB).

9.16.1.2.2 Birbhum Coalfield Area, West Bengal

Coal exploration, by detecting Gondwana sediments, is also a proven application in India (Naskar and Lahiri, 2014). Magnetotelluric surveys in the eastern part of the Birbhum coalfield area, West Bengal, were conducted for imaging crustal electrical conductivity, including the thickness of Rajmahal trap and the Gondwana sediments, for better understanding of the conceptual model of tectonic setup and to identify new and larger target area. 1-D inversion model of MT data indicates a highly conductive layer of resistivity (1–7 $\Omega\text{-m}$) underlying the Rajmahal trap with a varying thickness of 522–910 m. This intermediate conductive layer is interpreted as Gondwana sediments with a maximum thickness of 910 m at Sherpur and Nabipur, respectively (Naskar and Lahiri, 2014). Subsequently, a high-resistive subsurface of resistivity more than 500 $\Omega\text{-m}$ is interpreted beneath the conductive Gondwana layer. This resistive interface could be attributed to the Archean basement lying at a varying depth of 1,232–1,790 m along the profile. A steep gradient in the basement is observed near the station Purapara which could be attributed to fault. This area also coincides with the steep gravity gradient zone indicated in the Bouguer gravity anomaly map (Saha et al., 1992) that is attributed to basement fault. The basement appears to slope down in the east towards the Bengal geosyncline.

9.16.1.3 Geothermal Exploration

MT geothermal exploration measurements allow detection of resistivity anomalies associated with productive geothermal structures, including faults and the presence of a cap rock, and allow for estimation of geothermal reservoir temperatures at various depths.^{12, 13, 14} Dozens of MT geothermal exploration surveys have been completed in Japan and the Philippines since the early 1980s, helping identify several hundred megawatts of renewable power at places such as the Hatchobaru plant on Kyushu^{15, 16} and the Togonang plant on Leyte.^{17, 18, 19} Geothermal exploration with MT has also been done in the United States, Iceland,²⁰ New Zealand, Hungary, China,²¹ Ethiopia, Indonesia, Peru,²² Australia, and India (Naskar, 2017; Israil et al., 2020). Following are some of the links related to the application of MT to geothermal exploration for detailed information.

9.16.1.3.1 Chabsar Area, Ahmedabad, Gujarat

Geothermal source characterization near the Chabsar hot water spring site, Ahmedabad, was carried out on the basis of the resistivity model obtained from MT profiles (Israil et al., 2020, profile ‘6’). The inverted resistivity model, in the form of depth slices, is shown in the figure in Israil et al. (2020). A low-resistivity (<10 $\Omega\text{-m}$) zone is found in the depth range between 1 and 2.5 km. A sharp contrast of resistivity divides the area into two parts along the NW–SE strike associated with the deep-seated (1 km) West Cambay fault (WCF). The geothermal source zone is connected with the WCF, which acts as a conduit for heat transfer from deeper level to shallow level (Mohan et al., 2017).

9.16.1.4 Other Commercial Applications

MT is also used for groundwater exploration and mapping, hydrocarbon reservoir monitoring, deep investigation (100 km) of the electrical properties of the bedrock for high-voltage direct current (HVDC) transmission systems,²³ carbon dioxide sequestration,^{24, 25} and other environmental engineering applications (e.g. nuclear blast site monitoring²⁶ and nuclear waste disposal site monitoring). Following are some links that demonstrate the applications of MT in different commercial investigations.

9.16.2 RESEARCH APPLICATIONS

9.16.2.1 Crustal Research

MT is treated as the most reliable supportive tool to cross-check the geophysical anomalous zones deciphered by seismological studies. MT has been used to investigate the distribution of silicate melts in the Earth's mantle and crust; large investigations have focused on the continental US (National Science Foundation Earth Scope MT Program), the East Pacific Rise, and the Tibetan Plateau. Other research work aims to better understand the plate tectonic processes in the highly complex three-dimensional region formed by the collision of the African and European plates, as evident from the outcome presented in the research available in the following web link. MT survey has been used for deep crustal studies in Central India (Naskar et al., 2012) and over Sikkim–Darjeeling Himalaya (Naskar et al., 2011; Naskar and Saha, 2016).

9.16.2.1.1 *Along the Siliguri–Gangtok Profile, Sikkim–Darjeeling Himalaya*

Naskar and Das (2015) conducted MT survey over Sikkim–Darjeeling Himalaya along the Siliguri–Gangtok transect in order to prepare a better conceptual model of the Earth crust (Naskar and Das, 2015). The 2-D cross section shows a high-conductivity feature beneath the Chamakdangi–Tarkhola region extending from a shallow surface into the deep crust. Studies over subducting plates in several parts of the world have revealed similar high-conductivity layers (Jones, 1992; Jones et al., 1992; Adam, 1980). The geomagnetic depth sounding studies in the NW Himalaya show a similar high conductivity associated with the HFT (Reddy and Arora, 1993). Studies on the depth of the earthquake foci in this region (Ni and Barazangi, 1984) indicate that the top of the subducting Indian plate is about 10 km beneath the Siwalik region. The top (about 4 km) may be due to the Siwalik sediments thrust down along the MBT.

9.16.2.1.2 *Along the Bargaon–Barwah Profile across NSL, Central India, Madhya Pradesh (MP)*

The Narmada–Son lineament (NSL) in the CIS is a profound crustal discontinuity reactivated several times in the geologic past with left lateral transcurrent motion. These episodic reactivations have produced many graben (rifts) and horsted blocks (fold belts) starting from early Proterozoic, when the volcanosedimentary Mahakoshal Supergroup was deposited in a rifted basin. This was followed by the deposition of the Vindhyan Supergroup, the Gondwana Supergroup, the Lameta Group, the Tertiary Deccan basalts, the Quaternary/Recent alluvium, and the laterites. In the present study, the authors tried to re-examine the crustal structure in the Central Indian Shield (CIS) along the NSL by gravity, MT, and deep seismic sounding (DSS) data (Naskar et al., 2012). The TE and TM mode MT data was inverted using the rapid relaxation inversion (RRI) method of Smith and Booker (1991) to generate subsurface conductivity distribution in the area. Below stations 5, 6, 7, and 8, there is a conductive structure starting from a shallow depth to deep into the crust. Highly conducting rocks are found to extend from the lower crust to the uppermost mantle, which indicates anomalously high heat flow at depth and might be the source for the generation of several geothermal springs in this area, since the higher the temperature, the higher the conductivity of silicate rocks at depth. Rifting of the continental crust along the NSL as indicated by the large extent of

the conductive zone prominent in the model, and it has also been speculated that the continent has a deep-rooted rift along the NSL (Naskar et al., 2012) ('Picasso-Phase I: MT Investigation of the Betic-Rif mountain system. Comparison of actual robust processing algorithms'. [www.geophysics.dias.ie/Picasso/data/Schmoldt-Extended Abstract](http://www.geophysics.dias.ie/Picasso/data/Schmoldt-Extended%20Abstract)).

9.16.2.1.3 Along the Mundgod–Siruguppa Profile, Dharwar Craton, South India

Magnetotelluric (MT) is the more dominant method for the imaging of subsurface conductivity features in the lateral and vertical directions. This method is greatly responsive to the change in various parameters, like temperature and composition, interconnected conductors, fluids, partial melts, metals, etc. Therefore, MT is the prevailing method to identify lithospheric-scale structures (Dehkordi et al., 2019; Jones, 1999 for a review). In the study of Kusham et al. (2021), MT data were used to image the crustal and upper mantle structures in the Dharwar craton, South India, along the CC' profile (Kusham et al., 2021). The details of the previous geophysical and geological studies carried out in the study region are summarized in Naganjaneyulu and Santosh (2012), Kusham et al. (2018, 2019). In the present study, both 2-D and 3-D inversion are carried out for the given dataset (CC') (Kusham et al., 2021). Firstly, 2-D inversion (Kusham et al., 2021) was carried out to delineate the structures beneath the profile. A lot of differences are observed in this model (Kusham et al., 2021) compared to earlier 2-D models (Kusham et al., 2021, along AA' and BB' profiles) in terms of the low-resistive feature mapped in the lithospheric uppermost mantle region. From the 2-D inversion, no conductive feature is mapped in the deeper lithosphere, and the model shows a complete resistive nature (Kusham et al., 2021). In the earlier 2-D models, a conductor is present in the depth range of 100–250 km in the western Dharwar craton region (Kusham et al., 2021). Several slices in the crust and uppermost mantle at various depths—12 km, 15 km, 18 km, 40 km, 100 km, 140 km, 170 km, and 200 km—are shown in the figure in Kusham et al. (2021). The slices at different depths also show several crustal and uppermost mantle conductors. The figure in Kusham et al. (2021) shows the lithospheric magnetotelluric 3-D resistivity model with the Bouguer gravity anomaly curve (Subrahmanyam and Verma, 1982). The 2-D cross-section derived from the 3-D model (Kusham et al., 2021) defines a good representation of the most robust conductive features characterizing the crustal and uppermost mantle region throughout the profile. The 3-D inversion imaged several low resistive features in the crust (C1 and C2) and uppermost mantle (C3) region.

The 3-D inversion resistivity model contains prominent crustal conductors at the depth range of 0–40 km (Kusham et al., 2021). The features C1 and C2 are present in the shallow lithosphere and are associated with relatively low resistivity ($\sim 10 \Omega\text{-m}$), low gravity (-70 to -90 mGal; Subrahmanyam and Verma, 1982), and low heat flow values ($30\text{--}45 \text{ mW/m}^2$). The lateral variation observed on the Bouguer anomaly map compares well with the lateral distribution of the resistivity contrasts at depth. 3-D inversion results image a conductor in the depth range of 100–200 km (C3). The conductor having a bottom depth of about 200 km can be explained as due to the process of Cretaceous refertilization during the craton passage over several hotspots. Refertilization of the mantle can lead to long-lived enhanced conductivity. The conductor could be interpreted in the terms of the melt present in the region. A similar feature is also observed in the Precambrian Grenville Province margin in the Canadian Shield region (Adetunji et al., 2014). The 3-D inversion results are comparable to the earlier published 2-D inversion results of adjacent profiles (AA' and BB'). The concluding remark is that the 3-D image of the single profile data gives a better understanding of the subsurface structure as compared to 2-D inversion results.

9.16.2.1.4 Along the Chikmagalur–Kavali Profile, Dharwar Craton, South India

Broadband and long-period magnetotelluric measurements made at 63 locations along the ~ 500 km long ChikmagalurKavali profile that cut across the Dharwar Craton (DC) and Eastern Ghat

Mobile Belt (EGMB) in South India (Malleswari et al., 2019; Chandrakala et al., 2015). The 2-D resistivity model shows moderately conductive features (e.g. C1–C9) that intersperse a highly resistive background of crystalline rocks (e.g. R1–R3) and spatially connect to the exposed schist belts or granitic intrusions in the DC (Malleswari et al., 2019). These features are therefore interpreted as images of fossil pathways of the volcanic emplacements associated with the greenstone belt and granite suite formation exposed in the region. A near-vertical conductive feature (C8) in the upper mantle under the Chitradurga shear zone represents the Archean suture between the western and eastern blocks of DC. Although thick (~200 km) cratonic (highly resistive) lithosphere is preserved, a significant part of the cratonic lithosphere below the western DC is modified due to plume–continental lithosphere interactions during the Cretaceous–Tertiary period. A west-verging moderately conductive feature imaged beneath the EGMB lithosphere is interpreted as the remnant of the Proterozoic collision process between the Indian land mass and East Antarctica. Thin (~120 km) lithosphere is seen below the EGMB, which form the exterior margin of the Indian Shield subsequent to its separation from East Antarctica through rifting and opening of the Indian Ocean in the Cretaceous.

9.16.2.1.5 Along the Hamirpur–Rupadia Profile, Indo-Gangetic Plain

An electrical resistivity model along a 285 km long MT profile (Israil et al., 2020, profile ‘2’) extending between Hamirpur and Rupadia, in the Indo-Gangetic plain, is presented by Manglik et al. (2015). MT data was recorded at 39 sites along the profile in a NE–SW direction. The inverted resistivity depth model is shown in the figure in Israil et al. (2020). This model shows a high-resistive block in the southern end, which represents the Bundelkhand massif beneath 250–300 m thick sediments. The thickness of sediments gradually increases from south to north due to the Bundelkhand massif in the southern zone of the profile. The thickness of the sediment is 500–600 m at Kanpur, 1.2 km at Lucknow, increasing to 9 km at Behraich. The Lucknow fault is delineated by an abrupt increase in the sediment thickness to 2.5 km within a 20 km horizontal distance along the profile. The underlying rock is moderately resistive, representing the Vindhyan sequence.

9.16.2.1.6 Along the Tuwa–Godhra Profile, Gujarat

MT survey in the Tuwa–Godhra region (Gujarat) (Mohan et al., 2018b) is shown in the figure in Israil et al. (2020, profile ‘4’). The inverted resistivity depth model (Israil et al., 2020) shows a 500 m thick top layer with <700 Ωm resistivity. The second layer has a high resistivity (105 Ωm) and a thickness of 7 km in the central portion of the profile. Based on stratigraphic succession, this layer is interpreted as Godhra granite. The third layer has resistivity of 60–10,000 Ωm and is interpreted as Aravalli Supergroup. Godhra granite and Lunavada Group contact zone can be seen as a low-resistivity contrast between sites C13 and C15. The NW–SE striking Himmatnagar fault (HnF) is located in the central part of the profile, showing two prominent fault segments (sidestepped) near Tuwa. In between sites C3 and C12, the Aravallis might have been shifted downward (about 3.5 km) within the stepover zone of the HnF. An electrical conductance map of the Saurashtra region was presented by Kumar et al. (2018b).

9.16.2.1.7 The Narmada–Son Lineament (NSL) Zone, Central India, Madhya Pradesh (MP)

Patro and Sharma (2016) conducted MT investigations in the western segment of the Narmada–Son lineament (NSL) zone in Central India (Israil et al., 2020, profile ‘13’) and delineated the crustal resistivity structure (Israil et al., 2020). Locations of 153 sites MT data are shown on a simplified tectonic map in figure 5 of Israil et al. (2020). An inverted 3-D resistivity model in the form of vertical plain slices is shown in figure 6 of Israil et al. (2020). The conductive features were interpreted as mafic/ultramafic bodies derived from the mantle. These bodies represent the intrusive component

of the large igneous province (LIP) of the Deccan volcanic episode triggered by the passage of the Indian subcontinent over the reunion hotspot during the Late Cretaceous.

9.16.2.1.8 *Andaman–Nicobar Islands*

The Andaman–Nicobar ridge is situated at the northern segment of the Sunda subduction zone and forms the outer arc high in the forearc basin. The tectonic evolution of the Andaman–Nicobar ridge (Andaman–Nicobar Islands) occurred during the Cretaceous–Cenozoic period. The upliftment of the Andaman–Nicobar region occurred during the Eocene period. This region forms the outer arc accretionary sedimentary complex, and the inner igneous arc which runs parallel to the Andaman–Nicobar Islands is related to the Andaman–Sumatra active subduction zone. A long-period magnetotellurics (LMT) survey was conducted in the South Andaman, Middle Andaman, and Greater Nicobar regions (Patro et al., 2021, profile no. 11 for profile location) with the aim of mapping the deep geoelectric structures beneath this land mass, which may provide clues to understand the tectonic evolution of the Andaman–Nicobar region (Anusha et al., 2020). A forward response was computed from three LMT sites along each profile utilizing the seismic reflection modelling results based on seismic reflection data in this region, as shown in the figure in Patro et al. (2021). In the North Andaman, a moderately conductive body is delineated at crustal level inferred to be associated with the presence of ophiolites. In the Middle Andaman, a highly conductive structure is delineated which is interpreted to be due to the possible presence of fluids along the Diligent fault. In the Greater Nicobar region, the modelling study identified a conductive body related to the presence of a layered magmatic body, possibly linked to the volcanic seamount near the subducting plate (Anusha et al., 2020).

9.16.2.1.9 *Southern Part of the Deccan Volcanic Province (DVP), Maharashtra*

This study was carried out in the DVP region, which lies south of the NSL (Patro et al., 2021, profile no. 8). The DVP in the south is bounded by the Dharwar craton and Kaladgi and Bhima basins. To understand the lithospheric electrical structure in the DVP of the Indian Shield, MT studies were carried out along three traverses (Sangole–Partur [SP], Guhaghar–Koyna–Sangole [GKS], and Daulatabad–Koyna [DK]) (Patro and Sarma, 2009). The 2-D modelling results along the three profiles are shown in the figure in Patro et al. (2021). The thickness of the high-resistive lithosphere layer decreases from south to north. The MT model brought out an upper mantle conductor with a conductance of about 2,000 S extending over a length of about 100 km laterally and 80 to 120 km vertically. The upper mantle conductor may be the signature of the postulated subduction zone of the Western Dharwar Craton (WDC) underneath the Eastern Dharwar Craton (EDC). The MT result along the GKS profile, located at the southern edge of the DVP and passing through the Koyna region, shows a lithospheric electrical signature similar to that observed along other traverses (SP and DK) but suggests a gradual thinning towards the western half of the profile. Broadband MT data were acquired along a 35 km long profile in the Palghar region to understand the swarm-type earthquake activity. The 2-D geoelectric structure along the profile delineated a conductor and highly resistive basement beneath the Palghar region. MT modelling–derived conductor coincides with the microearthquakes, whereas the interaction zone between the conductor and resistivity formation acts as a zone of moderate earthquakes (Pavankumar et al., 2020).

9.16.2.1.10 *Southern Granulite Terrain (SGT)*

The Southern Granulite Terrain (SGT), located in the southern portion of the Indian peninsula, was formed during the Archean to the Proterozoic times and is considered as one of the largest granulite terrains of the globe. It consists of retrograded hard crystalline rocks, and the SGT is divided into three different blocks from north to south, viz., Salem, Madurai, and Trivandrum blocks (e.g. SB,

MB, and TB). The major structural features, the Palghat Cauvery shear zone (PCSZ), separates SGT and Dharwar Craton. Besides a large number of other geological and geophysical studies, magnetotelluric studies were also carried out in this region to decipher the crustal signature of PCSZ and to provide insights into crust and lithospheric electric signatures (Harinarayana et al., 2003, 2006a; Naganjaneyulu and Harinarayana, 2003; Naganjaneyulu and Santosh, 2010; Naidu et al., 2011; Patro et al., 2014). A 3-D geoelectric model of SGT derived using all the available MT data acquired until 2004 (Patro et al., 2021, profile no. 8 for location) is shown in the figure in Patro et al. (2021, 2014). It has delineated a well-defined major crustal conductor (C3) located in the northern half of the Madurai block. The electrical characteristics of this south-dipping conductor and its close spatial correlation with two of the major structural elements (R2 and R3), viz., the Karur–Oddanchatram–Kodaikanal shear zone and the Karur–Kamban–Painavu–Trichur shear zone, suggest that this conductive feature is closely linked to the subduction–collision tectonic processes in the SGT. It is inferred that the Neoproterozoic SGT terrain boundary lies south of the Palghat–Cauvery shear zone. The modelling results also showed that the Achankovil shear zone (ACSZ) is characterized by a well-defined north-dipping conductive feature (C4). The Trivandrum block (resistive feature R4) on the southern side is seen as downthrown relative to the Madurai block, suggesting a northward movement of the Trivandrum block colliding against the Madurai block. The northern lithospheric segment in SGT is characterized by a thick resistive upper mantle, while the southern one is characterized by a conductive medium, suggesting a relatively thinned lithosphere in the southern segment.

9.16.2.2 Earthquake Precursor Prediction Research

Fluctuations in the MT signal may be able to predict the onset of seismic events. Stationary MT monitoring systems have been installed in Japan since April 1996, providing a continuous recording of MT signals at the Wakuya Station (previously at the Mizusawa Geodetic Observatory) and the Esashi Station of the Geographical Survey Institute of Japan (GSIJ). These stations measure fluctuations in the Earth's electromagnetic field that correspond with seismic activity. The raw geophysical time series data from these monitoring stations is freely available to the scientific community, enabling further study of the interaction between electromagnetic events and earthquake activity. The MT time series data from the GSIJ earthquake monitoring stations is available online. Additional MT earthquake precursor monitoring stations in Japan are located in Kagoshima, in Sawauchi, and on Shikoku. Similar stations are also deployed in Taiwan on Penghu Island, as well as in the Fushan Reserve on the island of Taiwan proper. The presence of fluids at the 1993 Latur earthquake and the 2001 Bhuj source zone in the crust beneath the Indian stable continental plate interior inferred by seismic tomography (Gupta, 1993; Gupta et al., 1993; Mishra and Zhao, 2003) was found supported by the MT survey (Sarma et al., 1994; Arora et al., 2002).

POLARIS is a Canadian research program investigating the structure and dynamics of the Earth's lithosphere and the prediction of earthquake ground motion. MT survey has been carried out in the Shillong Plateau and Assam Valley area for deciphering the south-dipping hidden Oldham fault at the plateau and valley boundary (Saha et al., 2007).

9.16.2.2.1 *The Borgaon–Barwah Profile across the NSL, Central India, Madhya Pradesh (MP)*

Naskar et al. (2008) conducted an MT survey across the Narmada–Son lineament (NSL) in order to delineate the neotectonic activities when a swarm of earthquakes occurred at Pandhana in 1998. The MT apparent resistivity pseudosection shows the zones of weaknesses as closely spaced vertical contour lines (Figure 9.2). These are representatives of vertical planes of weaknesses, like faults, where electric charges accumulate. There are several zones of such vertical faults around stations 2, 4, 5, 6, and station 8. Vertical discontinuities at stations 5 and 6 are associated with phase shift at shallower depth; those at stations 2, 4, and 8 are not corroborated by the phase shift. Thus, the

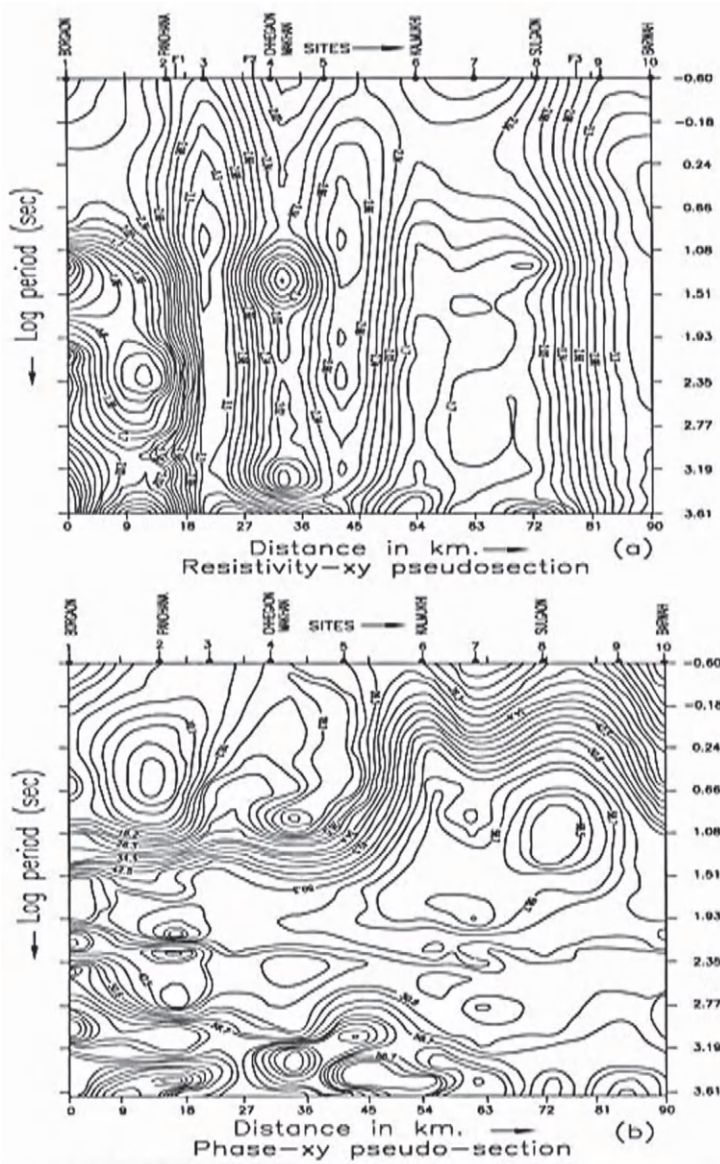


FIGURE 9.2 (a) Apparent resistivity and (b) phase pseudosections using TE mode MT data, Central India, MP.

Source: Naskar et al. (2008).

faults at stations 2, 4, and 8 are quite obvious. Special mention may be made of the fault that has been inferred from the MT survey at Pandhana where a swarm of earthquakes occurred during 1998. This indicates the active nature of the fault where neotectonic activity has been encountered.

- 'Terrestrial, Atmospheric and Oceanic Sciences' (<http://tao.cgu.org.tw/index.php?id=380>). Tao.cgu.org.tw. 21 September 1999. Retrieved 18 October 2011.
- 'Short-term earthquake prediction: Current status of seismo-electromagnetics'. Tectonophysics. 470: 205–213. Bibcode: 2009Tectp.470.205U (<http://adsabs.harvard.edu/abs/2009Tectp.470.205U>). doi:10.1016/j.tecto.2008.07.019 (<https://doi.org/10.1016%2Fj.tecto.2008.07.019>).

- 'A critical review of Electric Earthquake Precursors' (www.earth-prints.org/bitstream/2122/1213/1/22-1.pdf)(PDF).
- 'Sawauchi Automated Stationary MT Data and Earthquake Activity (>4.0M) during May ~ August, 2008' (www.phoenix-geophysics.com/applications/earthquake_research/short_report_sawauchi.pdf) (PDF). 2008.
- https://web.archive.org/web/20100225080738/http://vldb.gsi.go.jp/sokuchi/geomag/menu_03/mt_data-e.html
- <http://tao.cgu.org.tw/pdf/v133p367.pdf>
- 'Polaris Consortium' (www.polarisnet.ca/). Polarisnet.ca. Retrieved 18 October 2011.

9.16.2.2.2 Along the Kainad–Kinhwali Profile, Palghar Region, Maharashtra

The Deccan volcanic province (DVP) in the western part of peninsular India consists of a thick pile of flood basalts resting mainly on the Archaean and Proterozoic rocks forming the basement. This intraplate region experiences moderate seismic activity, the most recent one being a swarm-type activity in the Palghar region, about 120 km north of Mumbai, that started in November 2018 and has produced a few thousand microearthquakes and a 4.3 magnitude earthquake since then. Pavankumar et al. (2020) have carried out a magnetotelluric (MT) study along a 35 km long profile across the seismic zone to delineate the subsurface structure to understand the possible cause for the seismic activity. The final 2-D inversion resistivity model is shown in the figure in Pavankumar et al. (2020). The average elevation along a perpendicular distance of 2 km and regional Bouguer gravity anomaly (BGA) curve (GSI-NGRI, 2006; Mishra et al., 2008) along the profile are also shown in the figure. Surface location of the west coast fault (Biswas, 1987) and the axis of the Panvel flexure in the model are marked. The geoelectric structure yields an assemblage of highly resistive and moderately conductive blocks in the uppermost crust resting on a major listric-type fault that possibly reaches the surface at the west coast fault from a depth of about 15 km beneath the Panvel flexure zone. In order to understand the relation between the conductive/resistive features and seismicity, the hypocenters of the swarm activity published by Sharma et al. (2020) are superimposed. These events were recorded during 26 January–30 May 2019 by a network of six broadband seismological stations distributed around the seismogenic zone. One of these broadband stations (Dhundhalwadi, DDWD) is located within the zone of the swarm activity. In conjunction with the regional Bouguer gravity anomaly and the seismicity distribution, it is inferred that the upper crustal heterogeneities, coupled with the basement fault and low rheological strength of the fractured upper to middle crust, might be leading to triggering of the seismicity in the region.

Sharma et al. (2020) attributed the Palghar swarm activity to the aseismic slip driven by fluid migration at a shallow depth by correlating the seismicity with the rainfall data and the surface deformation estimated from the InSAR analysis. The present results showing the distribution of small microearthquakes over the entire area of the conductor C1 and moderate earthquakes coinciding with the interface between the conductor and the highly resistive basement and intrusive imply that both fluids and localization of tectonic stresses should have an important role in controlling the seismicity of the Palghar region. Multidisciplinary studies including 3-D MT modelling of the data would provide better understanding on the tectonic processes that are undergoing in the study region.

9.16.2.2.3 Western Part of Wagad Region, Gujarat

A 3-D MT study was conducted by Azeez et al. (2018). They recorded MT data at 37 sites, covering an approximate area of 65×16 km² in the western part of the Wagad uplift, Gujarat (Israil et al., 2020, profile '12'). Locations of MT sites are shown in the figure in Israil et al. (2020) along with the geology and tectonic features. The objective of this study was to delineate the crustal geoelectric structure and to understand the subsurface geometry of faults/lineaments. These provide critical information to understand the earthquake generation mechanism/aftershocks pattern in the region. Full impedance tensor data was inverted using the 3-D MT

inversion algorithm of Siripunvaraporn et al. (2005a). The 3-D resistivity model in the form of four vertical plain slices is shown in the figure in Israil et al. (2020). A thick (~3 km) low-resistivity (1–30 Ω -m) layer depicts the Tertiary/Mesozoic sediments. Below the upper sediments, heterogeneous upper and middle crusts, characterized by conductive (<30 Ω -m) and moderately resistive (>300 Ω -m) features, are delineated. Enhanced conductivity zones are interpreted as fluid-enriched zones in the crust and correlate with the surface trace of various faults/lineaments. These conductive zones play an important role in earthquake triggering. The results provide a strong clue to infer that the fluid enrichment in the Khadir transverse fault (KTF) and Manfara fault (MF) plays an important role in the observed seismic activity and controls the aftershocks pattern.

NOTES

- 1 www.kmstechnologies.com www.kmstechnologies.com/ (www.kmstechnologies.com/). Retrieved 26 September 2016. *Missing or empty |title= (help)*.
- 2 “Surveys | AMT and MT” (<https://web.archive.org/web/20111003125930/www.zonge.com/FieldAMT.html>). Zonge. *Archived from the Original* (www.zonge.com/FieldAMT.html) on 3 October 2011. Retrieved 18 October 2011.
- 3 “GEPARD-8” (<http://agcos.ca/wp/product/gepard-8/>). *agcos.ca*. Retrieved 28 October 2017.
- 4 “Phoenix products: The MTU Receiver” (www.phoenix-geophysics.com/products/receivers/mtu/). *Phoenix geophysics.com*. Retrieved 18 October 2011.
- 5 “LEMI LLC” (www.lemisensors.com/). LEMI LLC. Retrieved 26 September 2016.
- 6 “Metronix Geophysics: Technology for Geophysics and Magnetotellurics” (www.geo-metronix.de/). *Metronix*.
- 7 “Vega Geophysics Official Web Site” (<http://vega-geophysics.com/>). Retrieved 28 March 2012.
- 8 “3-D MT Survey in Uzbekistan” (www.phoenix-geophysics.com/news/newsletter/?&issue=47&article=427). Phoenix-geophysics.com. Retrieved 18 October 2011, and “CSIRO Publishing—ASEG Extended Abstracts” (www.publish.csiro.au/paper/ASEG2003ab054.htm). Publish.csiro.au. Retrieved 18 October 2011.
- 9 “System 2000 Fuels Exploration Boom” (www.phoenix-geophysics.com/news/newsletter/?&issue=19). Phoenix-geophysics.com. Retrieved 18 October 2011.
- 10 www.terrapub.co.jp/journals/EPS/pdf/2002/5405/54050575.pdf
- 11 “Imaging the Geometry and Structure of Kimberlite Pipes Using Audio-MT” (http://homepages.dias.ie/~ajones/PowerPoint_Presentations/kimberlite/sld001.htm). Homepages.dias.ie. Retrieved 18 October 2011.
- 12 “Geothermal Exploration with Electromagnetic Methods” (www.kmstechnologies.com/Files/MT_for_Geothermal_Exploration_2008_FINAL_10.17.08.pdf) (PDF). 2008. Retrieved 18 October 2011.
- 13 “Mapping Geothermal Reservoirs Using Broadband 2-D MT and Gravity Data” (www.kmstechnologies.com/Files/Mapping_geothermal_reservoirs_using_broadband_2-D_MT_and_gravity_data_01.pdf) (PDF).
- 14 “Characterizing a Geothermal Reservoir Using Broadband 2D MT Survey in Theistareykir, Iceland” (www.kmstechnologies.com/Files/Characterizing_a_geothermal_reservoir_using_broadband_2-D_MT_survey_01.pdf) (PDF). Retrieved 18 October 2011.
- 15 “Magnetotelluric Soundings in the Takigami Geothermal Area, Japan” (www.geothermal-energy.org/pdf/IGA_standard/WGC/2005/0718.pdf) (PDF). International Geothermal Association. 24 April 2005. Retrieved 24 January 2018.
- 16 “Science Links Japan | Geothermal Reservoirs Modeling in the Western Side of Mt. Aso, SW Japan by Magnetotelluric Method” (<https://web.archive.org/web/20120229033055/http://sciencelinks.jp/j-east/article/200512/000020051205A0468633.php>). Sciencelinks.jp. 18 March 2009. Archived from the original (<http://sciencelinks.jp/j-east/article/200512/000020051205A0468633.php>) on 29 February 2012. Retrieved 18 October 2011.
- 17 Josephine B. Rosell and Maribel C. Zaide-Delfin (24 April 2005). “Resource Potential of the Southern Leyte Geothermal Prospect, Philippines: A Geologic Evaluation” (www.geothermal-energy.org/pdf/IGAstandard/WGC/2005/0611.pdf) (PDF). International Geothermal Association. Retrieved 24 January 2018.
- 18 “Philippine National Oil Company” (<https://web.archive.org/web/20111002074359/www.pnoc.com.ph/about/milestones.php?page=3>). Pnoc.com.ph. Archived from the original (www.pnoc.com.ph/about/milestones.php?page=3) on 2 October 2011. Retrieved 18 October 2011.

- 19 “Geothermal | The Energy Development Corporation Website” (<https://web.archive.org/web/20151104154602/www.energy.com.ph/our-projects/geothermal/>). Energy.com.ph. Archived from the original (www.energy.com.ph/our-projects/geothermal/) on 4 November 2015. Retrieved 18 October 2011.
- 20 “Characterizing a Geothermal Reservoir Using Broadband 2-D MT Survey in Theistareikir, Iceland” SEG Expanded Abstracts. 2008.
- 21 www.bgp.com.cn/download.aspx?id=156
- 22 “Geothermal Mt Survey in Peru” (www.phoenix-geophysics.com/news/newsletter/?issue=44). Phoenix geophysics.com. Retrieved 18 October 2011.
- 23 H. Thunehed; et al. “Geophysical and Geological Pre-Investigations for HVDC-Electrodes” (<http://ieeexplore.ieee.org/iel5/4492962/4498028/04498123.pdf>) (PDF). PowerAfrica 2007. IEEE, DOI: 10.1109/PESAFR.2007.4498123 (<https://doi.org/10.1109/2FPESAFR.2007.4498123>).
- 24 “Energy Sector: Science and Technology: Cleaner Fossil Fuels” (<https://web.archive.org/web/20110811170300/www.nrcan.gc.ca/eneene/science/tirdrdti/clepro-eng.php>). Natural Resources Canada. 4 May 2010. Archived from the original (www.nrcan-rncan.gc.ca/eneene/science/tirdrdti/clepro-eng.php) on 11 August 2011. Retrieved 18 October 2011.
- 25 “MT Survey in Taiwan Evaluates the Possibility of CO₂ Sequestration” (www.phoenix-geophysics.com/news/newsletter/?&issue=43&article=397). Phoenix-geophysics.com. Retrieved 18 October 2011.
- 26 “Hydrogeologic Assessment of the Amchitka Island Nuclear Test Site (Alaska) with Magnetotellurics” (<http://cat.inist.fr/?aModele=afficheN&cpsidt=18787533>). Cat.inist.fr. Retrieved 18 October 2011.

REFERENCES

- Adam, A., 1980. The change of electrical structure between an orogenic and an ancient tectonic area. *Journal of Geomagnetism and Geoelectricity*, vol. 32, pp. 1–46.
- Adetunji, A.Q., Ferguson, I.J. and Jones, A.G., 2014. Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin. *Tectonophysics*, vol. 614, pp. 146–169.
- Ajithabh, K.S. and Patro, P.K., 2021. Crustal deformation in Volcanic covered area as inferred from magnetotelluric studies: An example from India. *Journal of Geodynamics*, Elsevier, Netherlands, vol. 145, p. 101840.
- Anusha, E., Arora, K., Nagarajan, N. and Azeez, K.A., 2020. Constraints on the configuration of Andaman-Nicobar Subduction zone from EM modeling. *Tectonophysics*, vol. 792, p. 228575.
- Arora, B.R., Rawat, G. and Singh, A.K., 2002. *Mid-Crustal Conductor below the Kutch Rift Basin and Its Seismogenic Rlevance to the 2001 Bhuj Earthquake*. New Delhi, India: DCS-DST News, Govt. of India, pp. 22–24.
- Azeez, K.A., Mohan, K., Veeraswamy, K., Rastogi, B.K., Gupta, A.K. and Harinarayana, T., 2018. 3D crustal resistivity structure beneath the Wagad aftershock zone of the 2001 Bhuj earthquake, Kutch, India: Heterogeneous resistivity structure controlled by widespread fluid infiltration and clues to aftershocks pattern. *Tectonophysics*, vol. 747, pp. 54–67.
- Berdichevsky, M.H., 1968. *Elektriceskaja razvedka metodom magnetotelluriceskogo profilirovanija (Electric survey using the method of magnetotelluric profiling)*. Moskwa: Nedra, pp. 1–253.
- Berdichevsky, M.N. and Dimitriev, V.I., 1976. Distortion of magnetic and electric fields by near surface lateral inhomogeneities. *Acta Geodaetica et Geophysica*, Hungarian Academy of Sciences, vol. 11, pp. 447–485.
- Biswas, S.K., 1987. Regional tectonic framework, structure and evolution of the western continental marginal basins of India. *Tectonophysics*, vol. 135, pp. 307–327.
- Cagniard, L., 1953. Basic theory of the magneto-telluric method of geophysical prospecting. *Geophysics*, vol. 18, no. 3, pp. 605–635.
- Caldwell, T.G., Bibby, H.M. and Brown, C., 2004. The magnetotelluric phase tensor. *Geophysical Journal International*, vol. 158, p. 457.
- Campbell, W.H., 1997. *Introduction to Geomagnetic Fields*. Cambridge: Cambridge University Press.
- Chandrakala, K., Pandey, O.P., Prasad, A.S.S.S.R.S. and Sain, K., 2015. Seismic imaging across the Eastern Ghats Belt-Cuddapah Basin collisional zone, southern Indian Shield and possible geodynamic implications. *Precambrian Research*, vol. 271, pp. 56–64.
- Dehkordi, B.H., Ferguson, I.J., Jones, A.G., Ledo, J. and Wennberg, G., 2019. Tectonics of the northern Canadian Cordillera imaged using modern magnetotelluric analysis. *Tectonophysics*, vol. 765, pp. 102–128.
- Eggers, D.E., 1982. An eigenstate formulation of the magnetotelluric impedance tensor. *Geophysics*, vol. 47, pp. 1204–1214.

- Everett, J.E. and Hyndman, R.D., 1967a. Magnetotelluric investigations in southwestern Australia. *Physics of Earth and Planetary Interiors*, Elsevier, vol. 1, pp. 49–54.
- Everett, J.E. and Hyndman, R.D., 1967b. Geomagnetic variations and electrical conductivity structure in southwestern Australia. *Physics of Earth and Planetary Interiors*, Elsevier, vol. 1, pp. 24–34.
- Feynman, R.P., Leighton, R.B. and Sands, M.W., 2005. *The Feynman Lectures on Physics: The Definitive and Extended Edition*. Boston, MA: American Publisher, Addison Wesley.
- Gamble, T.D., Goubau, W.M. and Clarke, J., 1979. Magnetotellurics with a remote magnetic reference. *Geophysics*, vol. 44, pp. 53–68.
- Gamble, T.D., Goubau, W.M. and Clarke, J., 1979. Magnetotellurics with a remote magnetic reference. *Geophysics*, vol. 44(1), pp. 53–68. <https://doi.org/10.1190/1.1437915>.
- Groom, R.W. and Bailey, R.C., 1989. Decomposition of magnetotellurics impedance tensor in the presence of local three dimensional galvanic distortions. *Journal of Geophysical Research, American Geophysical Union (AGU)*, U.S., vol. 94, pp. 1913–1925.
- GSI-NGRI, 2006. Gravity Anomaly Map of India on 1:2 Million Scale; Geological Survey of India, Hyderabad and National Geophysical Research Institute, Hyderabad, India.
- Gupta, H.K., 1993. The deadly Latur earthquake. *Science*, vol. 262, pp. 1666–1667.
- Gupta, H.K., Mohan, I., Rastogi, B.K., Rao, M.N. and Ramakrishna Rao, C.V., 1993. A quick look at the Latur earthquake of 30 September, 1993. *Current Science*, India, vol. 65, no. 7, pp. 517–520.
- Gupta, M.L., Sharma, S.R. and Sundar, A., 1991. Heat flow and heat generation in the Archaean Dharwar cratons and implications for the Southern Indian shield geotherm and lithospheric thickness. *Tectonophysics*, vol. 194, pp. 107–122.
- Harinarayana, T., Manoj, Naganjaneyulu, K., Patro, B.P.K., Begam, D., Murthy, D.N., Rao, M., Kumaraswamy, V.T.C. and Virupakshi, G., 2003. Magnetotelluric investigation along Kuppam—Palani Geotransect, South India 2D Modeling results. In: Ramakrishnan, M. (ed.), *Tectonic of Southern Granulite Terrain, Kuppam-Palani Geotransect. Memoirs of the Geological Society of India*, vol. 50, pp. 107–124.
- Harinarayana, T., Naganjaneyulu, K. and Patro, B.P.K., 2006a. Detection of a collision zone in south Indian shield region from magnetotelluric studies. *Gondwana Research*, vol. 10, no. 1–2, pp. 48–56.
- Heise, W., Caldwell, T.G., Bibby, H.M. and Brown, C., 2006. Anisotropy and phase splits in magnetotellurics. *Physics of the Earth and Planetary Interiors*, vol. 158, pp. 107–121.
- Israil, M., Kumar, D., Devi, A. and Gupta, P., 2020. Magnetotelluric investigations in India during last five years. *Proceedings of the National Academy of Sciences, India*, vol. 86, no. 1, pp. 553–568, DOI: 10.16943/ptinsa/2020/49793.
- Jiracek, G.R. (1990). Near-surface and topographic distortion in electromagnetic induction. *Surveys in Geophysics*, vol. 11, pp. 163–203.
- Jones, A.G., 1992. Electrical conductivity of the continental lower crust. In: Fountain, D.M., Arculus, R. and Kay, R.W. (eds), *Developments in Geotectonics, 23, Continental Lower Crust*. Amsterdam: Elsevier Science Publishers, pp. 87–143.
- Jones, A.G., 1999. Imaging the continental upper mantle using electromagnetic methods. *Lithos*, Elsevier, Netherlands, vol. 48, pp. 57–80.
- Jones, A.G., Cough, D.I., Kurtz, R.D., Delaurier, J.M., Boerner, D.E., Craven, J.A., Ellis, R.G. and McNeice, G.W., 1992. Electromagnetic images of regional structure in the southern Cordillera. *Geophysical Research Letters*, vol. 12, pp. 2373–2376.
- Jones, F.W. and Price, A.T., 1970. The perturbations of alternating geomagnetic fields by conductivity anomalies. *Geophysical Journal of the Royal Astronomical Society*, vol. 20, pp. 317–334.
- Jones, F.W. and Vozoff, K., 1978. The calculation of magnetotelluric quantities for three-dimensional inhomogeneities. *Geophysics*, vol. 43, pp. 1167–1175.
- Jupp, D.L.B. and Vozoff, K., 1976. Stable iterative methods for the inversion of geophysical data. *Geophysical Journal of the Royal Astronomical Society*, vol. 42, pp. 957–976.
- Kao, S.C. and Rankin, D.M., 1977. Magnetotelluric impedance tensor estimation: A comparison of methods. *Journal of Geophysical Research*, vol. 82(B10), pp. 1783–1792.
- Konda, S., Patro, P.K. and Rao, C.K., 2015. Geoelectric signatures of Palnad sub basin and Nallamalai Fold Belt, Cuddapah Basin, India. *Journal of the Geological Society of India*, vol. 86, pp. 377–382.
- Kumar, P.V., Rao, P.B.V., Singh, A.K. and Rao, C.K., 2018b. Electrical conductance map for Saurashtra region, Gujarat, India. *Current Science*, vol. 114, pp. 2175–2181.
- Kusham, P.A., Naick, B.P. and Naganjaneyulu, K., 2018. Lithospheric architecture in the Archaean Dharwar craton, India: A magnetotelluric model. *Journal of Asian Earth Sciences*, vol. 163, pp. 43–53.

- Kusham, P.A., Naick, B.P. and Naganjaneyulu, K., 2019. Crustal and lithospheric mantle conductivity structure in the Dharwar craton, India. *Journal of Asian Earth Sciences*, vol. 176, pp. 253–263.
- Kusham, N.B.P., Pratap, A. and Naganjaneyulu, K., 2021. 2-D versus 3-D magnetotelluric data interpretation: A case study from the Dharwar craton, India. *Tectonophysics*, vol. 816, p. 229028.
- Lilley, F.E.M., 1976. Diagrams for magnetotelluric data. *Geophysics*, vol. 41, no. 4, pp. 766–770. Australian National University, Canberra A.C.T. 2600, Australia, DOI: 10.1190/1.1440648.
- Lilley, F.E.M., 1993. Magnetotelluric analysis using Mohr circles. *Geophysics*, vol. 58, pp. 1498–1506.
- Madden, T.R. and Swift, C.M. Jr, 1969. Magnetotelluric studies of the electrical conductivity structure of the crust and upper mantle: In the earth's crust and upper mantle. In: Hart, P.J. (ed), *American Geophysical Union (AGU) Monograph*. Washington, DC: AGU, pp. 469–479.
- Malleswari, D., Veeraswamy, K., Azeez, K.A., Gupta, A.K., Babu, N., Patro, P.K. and Harinarayana, T., 2019. Magnetotelluric investigation of lithospheric electrical structure beneath the Dharwar Craton in south India: Evidence for mantle suture and plume-continental interaction. *Geoscience Frontiers*, Beijing, China, vol. 10, no. 5, pp. 1915–1930.
- Manglik, A., Adilakshmi, L., Suresh, M. and Thiagarajan, S., 2015. Thick sedimentary sequence around Bahraich in the northern part of the Central Ganga foreland basin. *Tectonophysics*, vol. 653, pp. 33–40.
- Mishra, D.C., Tiwari, V.M. and Singh, B., 2008. Gravity studies in India and their Geological Significance (Eds) Harsh Gupta and Fareeduddin, Golden Jubilee Memoir Geol. Soc. India. *Journal of Earth System Sciences*, vol. 66, pp. 329–372.
- Mishra, O.P. and Zhao, D., 2003. Crack density, saturation rate and porosity at the 2001 Bhuj, India, earthquake hypocenter: A fluid-driven earthquake? *Earth and Planetary Science Letters*, vol. 212, no. 3–4, pp. 393–405, DOI: 10.1016/S0012-821X(03)00285-1.
- Mohan, K., Chaudhary, P., Kumar, G.P., Kothiyari, G.C., Choudhary, V., Nagar, M., Patel, P., Gandhi, D., Kushwaha, D. and Rastogi, B.K., (2018b). Magnetotelluric investigations in Tuwa-Godhra Region, Gujarat (India). *Pure and Applied Geophysics*, vol. 175, pp. 3569–3589.
- Mohan, K., Chaudhary, P., Patel, P., Chaudhary, B.S. and Chopra, S., (2018a). Magnetotelluric study to characterize Kachchh Mainland Fault (KMF) and Katrol Hill Fault (KHF) in the western part of Kachchh region of Gujarat, India. *Tectonophysics*, vol. 726, pp. 43–61.
- Mohan, K., Kumar, G.P., Chaudhary, P., Choudhary, V.K., Nagar, M., Khuswaha, D., Patel, P., Gandhi, D. and Rastogi, B.K., 2017. Magnetotelluric investigations to identify geothermal source zone near Chabsar hotwater spring site, Ahmedabad, Gujarat, Northwest India. *Geothermics*, vol. 65, pp. 198–209.
- Naganjaneyulu, K. and Harinarayana, T., 2003. Evidence for continent-continent-collision zone in the South Indian Shield region. *Gondwana Research*, vol. 6, no. 4, pp. 902–911.
- Naganjaneyulu, K. and Santosh, M., 2010. The central India tectonic zone: A geophysical perspective on continental amalgamation along a Mesoproterozoic suture. *Gondwana Research*, vol. 18, no. 4, pp. 547–564.
- Naganjaneyulu, K. and Santosh, M., 2012. The nature and thickness of lithosphere beneath the Archean Dharwar Craton, southern India: A magnetotelluric model. *Journal of Asian Earth Sciences*, Elsevier, U.K., vol. 49, pp. 349–361.
- Naidu, G.D., Manoj, C., Patro, P.K., Sreedhar, S.V. and Harinarayana, T., 2011. Deep electrical signatures across the Achankovil shear zone, Southern Granulite Terrain inferred from magnetotellurics. *Gondwana Research*, vol. 20, no. (2–3), pp. 405–426.
- Naskar, D.C., 2017a. Magnetotelluric (MT) and resistivity survey along Belsara-rasunpur traverse delineating conductivity structure over Bakreswar geothermal region in Birbhum District, West Bengal. *Universal Journal of Geoscience*, vol. 5, no. 3, pp. 49–64, DOI: 10.13189/ujg.2017.050302.
- Naskar, D.C., 2017b. Geophysical surveys over Bakreswar geothermal region in Birbhum district, West Bengal. *Journal of Indian Geophysical Union*, Hyderabad, vol. 21, no. 5, pp. 343–352.
- Naskar, D.C. and Das, P.C., 2015. MT Survey over Sikkim-Darjeeling Himalaya-A Case Study. *Journal of Geophysics*, vol. 36, no. 1, pp. 13–29. A Quarterly Publication, ISSN 2230-9497, Regd.No.32238/80.
- Naskar, D.C., Das, L.K., Roy, K.K., Majumdar, R.K. and Choudhury, K., 2012. A re-examination of the crustal structure across Narmada-Son-Lineament (NSL), Madhya Pradesh, Central India using magnetotelluric and gravity modelling. *Journal of Geophysics*, vol. 33, no. 3, pp. 69–77.
- Naskar, D.C., Das, L.K., Roy, K.K. and Singh, O.P., 2008. Structural Studies of Crust and Upper Mantle in Parts of Madhya Pradesh, Central India. Geological Survey of India, Pan-African Event: India & Antarctica, Spl. Publ. no. 91, pp. 211–228.

- Naskar, D.C., Saha, D.K. and Das, L.K., 2011. Electrical imaging of shallow and deep crust using magnetotelluric (MT) survey along Islampur-Siliguri-10 mile transect, Sikkim-Darjeeling Himalayan region. *Indian Journal of Geosciences*, vol. 65, no. 1, pp. 45–56.
- Naskar, D.C. and Lahiri, A.K., 2014. Electrical imaging of shallow and deep crustal structure using magnetotelluric method in the eastern part of Birbhum Coalfield, West Bengal, India. *Journal of Geophysics*, vol. 35, no. 4, pp. 147–157.
- Naskar, D.C. and Saha, D.K., 2016. Electrical imaging of crustal structures by magnetotelluric (MT) and resistivity surveys along a traverse from Islampur to Karandighi in Bengal basin. *Journal of Geophysics*, vol. 37, no. 4, pp. 197–219.
- Ni, J. and Barazangi, M., 1984. Seismotectonics of the Himalayan collision zone: Geometry of the underthrusting Indian plate beneath the Himalaya. *Journal Geophysical Research*, vol. 89, no. B2, pp. 1147–1163.
- Parker, R.L., 1980. The inverse problem of electromagnetic induction: Existence and construction of solutions based upon incomplete data. *Journal Geophysical Research*, U.S., vol. 85, pp. 4421–4425.
- Parkinson, W., 1959. Directions of rapid geomagnetic variations. *Geophysical Journal of the Royal Astronomical Society*, Oxford University Press, vol. 2, pp. 1–14.
- Patro, P.K., 2017. Magnetotelluric studies for hydrocarbon and geothermal resources: Examples from the Asian region. *Surveys in Geophysics*, vol. 38, no. 5, pp. 1005–1041.
- Patro, P.K. and Sarma, S.V.S., 2009. Lithospheric electrical imaging of the Deccan trap covered region of western India. *Journal Geophysical Research*, United States, vol. 114, p. B01102.
- Patro, P.K. and Sarma, S.V.S., 2016. Evidence for an extensive intrusive component of the Deccan Large Igneous Province in the Narmada Son Lineament region, India from three dimensional magnetotelluric studies. *Earth and Planetary Science Letters*, vol. 451, pp. 168–176.
- Patro, P.K., Sarma, S.V.S. and Naganjaneyulu, K., 2014. Three dimensional lithospheric electrical structure of Southern Granulite Terrain, India and its tectonic implications. *Journal Geophysical Research*, United States, vol. 119, no. 1, pp. 71–82.
- Patro, P.K., Vijaya Kumar, P.V. and Sarma, S.V.S., 2021. MT/LMT studies for crust and upper mantle structure of India and its adjoining regions: Contribution of CSIR-NGRI. *Journal of Geological Society of India*, vol. 97, pp. 1251–1259.
- Pavankumar, G., Chakravarthi, N.N., Demudu Babu, M. and Manglik, A., 2020. Magnetotelluric study of an intraplate seismic zone in the Palghar region of the Deccan Volcanic Province, India. *Journal of Earth System Sciences*, vol. 129, p. 133.
- Pek, J. and Santos, F.A.M., 2006. Magnetotelluric inversion for anisotropic conductivities in layered media. *Physics of the Earth and Planetary Interiors*, vol. 158, pp. 139–158.
- Radhakrishnamurthy, C. and Likhite, S.D., 1970. Hopkinson effect, blocking temperature and Curie point in basalts. *Earth and Planetary Sciences Letters*, Elsevier, Amsterdam, Netherlands, vol. 7, pp. 389–396.
- Reddy, C.D. and Arora, B.R., 1993. Quantitative interpretation of geomagnetic induction response across the thrust zones of Himalaya along the ganga Yamuna valley. *Journal of Geomagnetism and Geoelectricity*, vol. 45, pp. 775–785.
- Rodi, W. and Mackie, R.L., 2001. Nonlinear conjugate gradients algorithm for 2-D magnetotelluric inversions. *Geophysics*, vol. 66, pp. 174–187.
- Roy, K.K., Bhattacharya, B.B., Singh, A.K. and Srivastava, S., 1998. Magnetotelluric survey in some parts of the Archean craton of Eastern India. *Indian Journal of Geology* (Special Issue-Naha Volume), vol. 70, pp. 157–162.
- Saha, D.K., Naskar, D.C., Bhattacharya, P.M. and Kayal, J.R., 2007. Geophysical investigation for the hidden Oldham fault: Magnetotelluric and Deep resistivity surveys in Shillong Plateau and Assam valley. *Geological Society of India*, vol. 69, pp. 359–372.
- Saha, S.N., Roy, A.K., Brahman, C.V., Sastry, C.B.K. and De, M.K., 1992. Geophysical exploration for coal-bearing Gondwana basins in the states of West Bengal and Bihar in northeast India. *Tectonophysics*, vol. 212, pp. 173–192.
- Sarma, S.V.S., Virupakshi, G., Harinarayana, T., Murthy, D.N., Prabhakar, S., Rao, E., Veeraswamy, K., Madhusudhana, R., Sarma, M.V.C. and Gupta, K.R.B., 1994. A wide band magnetotelluric study of the Latur earthquake region, Maharashtra, India. *Memoirs of the Geological Society of India*, vol. 35, pp. 101–118.

- Sarma, S.V.S., Virupakshi, G., Harinarayana, T., Murthy, D.N., Someswara Rao, M., Sastry, R.S., Nagarajan, N., Sastry, T.S., Sarma, M.V.C., Rao, M., Veeraswamy, K., Rao, S.P.E., Gupta, K.R.B., Lingaiah, A., Sreenivasulu, T., Raju, A.V.S.N., Patro, B.P.K., Manoj, C., Bansal, A., Kumaraswamy, V.T.C., Sannasi, S.R., Stephen, C. and Naganjaneyulu, K., 1998. *Integrated Geophysical Studies for Hydrocarbon Exploration. Saurashtra, India*. National Geophysical Research Institute-98-EXP-237.
- Sarma, S.V.S., Virupakshi, G., Murthy, D.N., Harinarayana, T., Sastry, T.S., Rao, M.S., Nagarajan, N., Veeraswamy, K., Sarma, M.S., Rao S.P.E. and Gupta K.R.B., 1992. Magnetotelluric Studies for Oil Exploration Over Deccan Traps, Saurashtra, Gujarat, India. National Geophysical Research Institute-92-LITHOS-125.
- Sasaki, Y., 2004. Three-dimensional inversion of static-shifted magnetotelluric data. *Earth Planets Space*, vol. 56, no. 2, pp. 39–248.
- Sharma, V., Wadhawan, M., Rana, N., Sreejith, K.M., Agrawal, R., Kamra, C., Hosalikar, K.S., Narkhede, K.V., Suresh, G. and Gahalaut, V.K., 2020. A long duration non-volcanic earthquake sequence in the stable continental region of India: The Palghar swarm. *Tectonophysics*, vol. 779, p. 228376.
- Simpson, F. and Bahr, K., 2005. *Practical Magnetotellurics*. Cambridge: Cambridge University Press. Available at www.cambridge.org/catalogue/catalogue.asp?isbn=0521817277.
- Siripunvaraporn, W. and Egbert, G., 2000. An efficient data-subspace inversion method for 2-D magnetotelluric data. *Geophysics*, vol. 65, no. 3, pp. 791–803.
- Siripunvaraporn, W., Egbert, G., Lenbury, Y. and Uyeshima, M., 2005a. Three dimensional magnetotelluric inversion: Data space method. *Physics of the Earth and Planetary Interiors*, vol. 150, pp. 3–14.
- Smith, J.T. and Booker, J.R., 1991. Rapid inversion of two and three dimensional magnetotelluric data. *Journal Geophysical Research*, vol. 96, pp. 3905–3922.
- Sims, P., 1971. Least squares estimation of magnetotelluric impedance tensors. *Geophysical Journal International*, vol. 15(3), pp. 289–302.
- Subrahmanyam, C. and Verma, R.K., 1982. Gravity interpretation of the Dharwar greenstone-gneiss-granite terrain in the southern Indian shield and its geological implications. *Tectonophysics*, vol. 84, pp. 225–245.
- Swift (Jr.), C.M., 1967. A magnetotelluric investigation of an electrical conductivity anomaly in the Southern United States. Ph.D. Thesis. Department of Geology and Geophysics, M.I.T, MA.
- Szarka, L. and Menville, M., 1997. Analysis of rotational invariants of the magnetotelluric impedance tensor. *Geophysical Journal International*, vol. 129, pp. 133–142.
- Tikhonov, A.N., 1950. In 1953 on determining electrical characteristics of the deep layers of the Earth's crust. *Doklady*, vol. 73, pp. 295–297.
- Tikhonov, A.N. and Arsenin, V.Y., 1977. *Solution to III-Posed Problems*. New York: John Wiley.
- Vozoff, K., 1991. The magnetotelluric method. In: Nabighian, M.N. (ed), *Electromagnetic Methods in Applied Geophysics: Applications*, vol. 2. Tulsa: Society of Exploration Geophysicists, pp. 641–712.
- Wameyo, P.M., 2005. Appendices to the report: Magnetotelluric and Transient Electromagnetic methods in geothermal prospecting, with examples from Menengai, Kenya. United Nations University-Geothermal Training Program, Iceland, Report 21 appendices, p. 26.
- Weidelt, P., 1999. 3-D conducting models: Implications of electrical anisotropy. In: *Three Dimensional Electromagnetics*. Tulsa: Society of Exploration Geophysicist, pp. 119–137.
- Wiese, H., 1962. Geomagnetische tiefensondierung. Teil II: Die streichrichtung der Untergrundstrukturen des elektrischen Widerstandes, erschlossen aus geomagnetischen variationen. *Journal of Pure and Applied Geophysics*, vol. 52, pp. 83–103.
- Yee, S. and Paulson, K.V., 1987. The canonical decomposition and its relationship to other forms of magnetotelluric impedance tensor analysis. *Journal of Geophysics*, vol. 61, pp. 173–189.
- Zhang, P., Pedersen, L.B., Mareschal, M. and Chouteau, M., 1993. Channelling contribution to tipper vectors: A magnetic equivalent to electrical distortion. *Geophysical Journal International*, vol. 113, pp. 693–700.

10 Seismic Method and Seismic Tomography

10.1 INTRODUCTION

In seismic surveying, seismic waves are created by controlled sources and propagate through the subsurface. These waves will return to the surface after reflection or refraction at geological boundaries. Instruments distributed along the surface detect the ground motion caused by these returning waves and measure the arrival times of the waves at different ranges from the source. These travel times will be converted to depth values. Subsurface geological interfaces can be systematically mapped. Seismic methods represent a natural development of the already long-established methods of earthquake seismology. In earthquake seismology, the travel times of earthquake waves are recorded at seismological observatories. They provide information on the gross internal layering of the Earth. In the same way, but on a smaller scale, seismic surveying provides a detailed picture of subsurface geology. Artificial sources, such as explosions, are used in the seismic method. Location, timing, and source characteristics are, unlike earthquakes, under the direct control of geophysicists.

Seismic methods represent the single most important geophysical method in terms of the amount of survey activity and the very wide range of its applications. The method is particularly well suited to the mapping of layered sedimentaries and is therefore widely used in the search for oil and gas. The method is also used for the mapping of near-surface sedimentary layers, the location of the water table, engineering application, and the determination of the depth to bedrock. Seismic surveying can be carried on at land or at sea.

Since the 1920s, seismic reflection techniques have been used to search for petroleum, and refraction techniques have been used in engineering applications. Additionally, since the 1980s, significant strides have been made in both near-surface seismic reflection surveying and in the development of shallow seismic refraction methods. Near-surface methods use an adaptation of parameters to high-resolution information—that is, the capacity to discriminate layers—and may provide results from layers that are 500 m in depth, such as deep-buried valleys. The main references for this section are Pelton (2005), Steeples (2005), and various chapters on seismic methods in Knödel et al. (1997), Rabbel (2006), and Yilmaz (2001).

10.2 SEISMIC WAVES

To understand the different types of seismic waves that can propagate through the ground away from a seismic source, some elementary concepts of stress and strain need to be considered.

10.2.1 STRESS AND STRAIN

An external force F is applied across an area A of the surface of a body. The ratio of force to area ($\frac{F}{A}$) is known as ‘stress’. Stress can be resolved into two components: (1) one at right angles to the surface (normal or dilatational stress), and (2) one in the plane of the surface (shear stress). A body subjected to stress undergoes a change of shape and/or size known as ‘strain’. According to Hooke’s law, stress and strain are linearly dependent, and the body behaves elastically until the yielding point is reached. This elastic strain is reversible so that the removal of stress leads to the removal of strain. At stresses beyond the yield point, the body behaves in a plastic or ductile manner and

permanent damage results. If further stress is applied, the body is strained until it fractures (Kearey and Brooks, 2002). Earthquakes occur when rocks are strained until fracture. In exploration seismology, the amounts of stress and strain lie well within the elastic behaviour of natural materials.

The linear relationship between stress and strain in the elastic field is specified by its various elastic moduli:

1. Young modulus E (Kearey and Brooks, 2002)

$$E = \frac{\text{Longitudinal stress} \left(\frac{F}{A} \right)}{\text{Longitudinal strain} \left(\frac{\Delta l}{l} \right)}$$

2. Bulk modulus K (Kearey and Brooks, 2002)

$$K = \frac{\text{Volume stress} (P)}{\text{Volume strain} \left(\frac{\Delta V}{V} \right)}$$

3. Shear modulus μ (Kearey and Brooks, 2002)

$$\mu = \frac{\text{Shear stress } \tau}{\text{Shear strain } \tan \theta}$$

4. Axial modulus ψ (Kearey and Brooks, 2002)

$$\psi = \frac{\text{Longitudinal stress} \left(\frac{F}{A} \right)}{\text{Longitudinal strain} \left(\frac{\Delta l}{l} \right)} \quad (\text{no lateral strain})$$

10.2.2 TYPES OF SEISMIC WAVES

Seismic waves are parcels of elastic strain energy that propagate from a seismic source, such as an earthquake or an explosion. The strains associated with the passage of a seismic pulse may be assumed to be elastic (except in the vicinity of the source). The propagation of seismic pulses is determined by the elastic moduli and densities of the materials through which they pass. There are two groups of seismic waves: body waves and surface waves.

10.2.2.1 Body Waves

Seismic waves include body waves that travel three-dimensionally through solid Earth volumes (Figure 10.1). For seismic investigation of the ground, the use of body waves is standard. Here we have to differentiate between compressional waves (P waves) and shear waves (S waves), and the difference is in the particle motion of the wave propagating through the underground material (Reynolds, 1998)

- P wave (*longitudinal, primary, or compressional wave*). Material particles oscillate about a fixed point in the direction of wave propagation by compressional and dilatational strain.
- S wave (*transverse, secondary, or shear wave*). Particle motion is at right angles to the direction of wave propagation and occurs by pure strain.

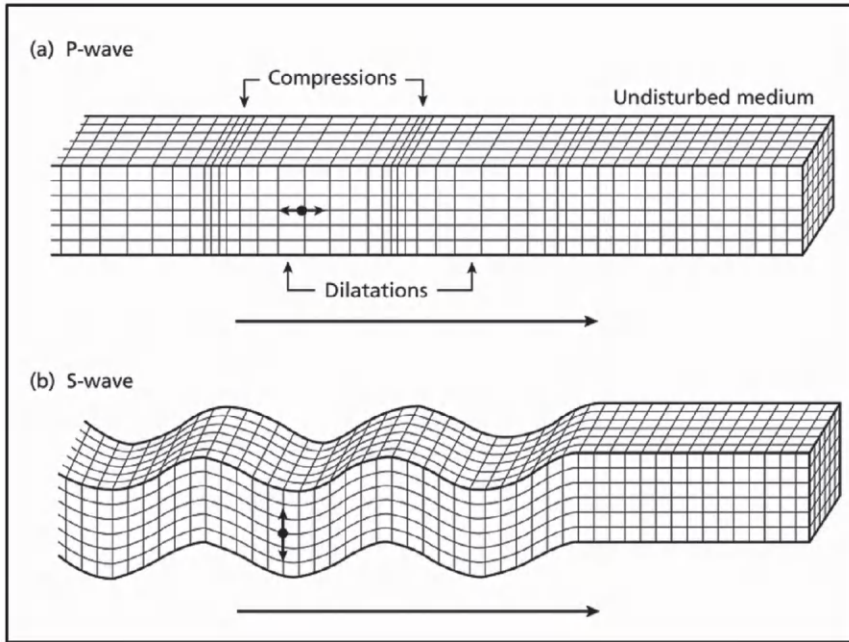


FIGURE 10.1 Elastic deformations and ground particle motions associated with the passage of body waves: (a) a *P* wave and (b) an *S* wave.

P waves have principally a higher velocity than *S* waves. *S* waves cannot propagate in fluids because fluids do not support shear stress.

10.2.2.2 Surface Waves

Waves that do not penetrate into the subsurface media are known as surface waves. Surface waves are categorized further as Love waves and Rayleigh waves (Figure 10.2); the ground roll, for example, is a Rayleigh wave (Reynolds, 1998).

- *Rayleigh wave*. Travels along the free surface of the Earth with amplitudes that decrease exponentially with depth; particle motion is, in an elliptical sense, in a vertical plane with respect to the surface. Rayleigh waves travel only through a solid medium.
- *Love wave*. Occurs only where a medium with a low *S* wave velocity overlies a layer with a higher *S* wave velocity; particle motion is at right angles to the wave propagation but parallel to the surface.

10.3 SEISMIC WAVE VELOCITIES

The velocities of seismic waves depend on the elasticity and density of the underground material and can be expressed by

$$V_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} = \sqrt{\frac{E}{\rho} \frac{1-\sigma}{2(1+\sigma)}} \quad (10.1)$$

$$\text{and } V_S = \sqrt{\frac{\mu}{\rho}} = \sqrt{\frac{E}{\rho 2(1+\sigma)}} \quad (10.2)$$

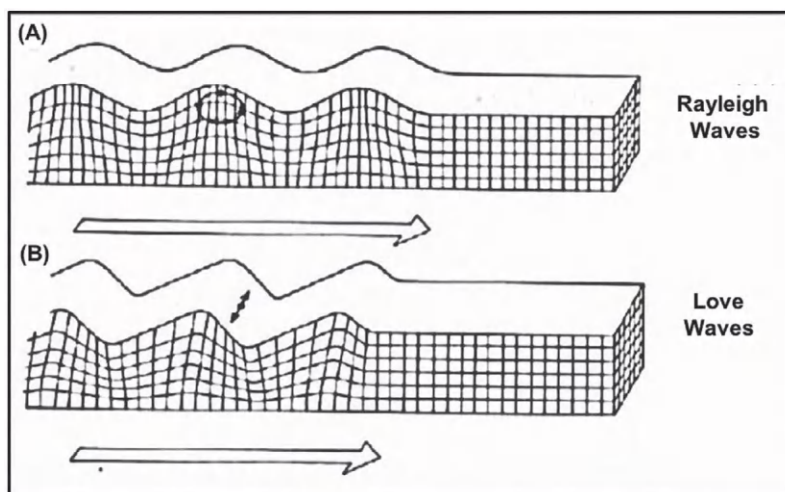


FIGURE 10.2 Elastic deformations and ground particle motion associated with the passage of surface waves: (a) a Rayleigh wave and (b) a Love wave.

with the elastic constants K (bulk modulus) $= \rho (V_p^2 - \frac{4}{3} V_s^2)$ and μ (shear modulus/rigidity modulus) $= \rho V_s^2$. V_p is the P wave velocity (m/s), V_s the S wave velocity (m/s), and σ Poisson's ratio.

10.3.1 ELASTIC WAVE VELOCITY AS A FUNCTION OF GEOLOGICAL AGE AND DEPTH

$$v = 1.47 \left(zT \right)^{\frac{1}{6}} \text{ km/s} \quad (10.3)$$

with z as depth in km, and T the geological age in millions of years.

Table 10.1 shows the range of values of propagation velocities, V_p and V_s , and the densities of principal rock types.

Normally, for unconsolidated sediments like gravel, sand, till, or clay, P wave velocities range from 200 to 800 m/s for dry and 1,500–2,500 m/s for water-saturated material. S wave velocities are much slower, in the range of 100–500 m/s, and do not differ between dry and water-saturated material. Due to its slowness, the S wave arrives later than the P wave and is named secondary (S) wave in contrast to the primary (P) wave. Shear waves do not propagate in fluids. The ratio σ (Poisson's ratio) $= \frac{V_p}{V_s}$ is independent of the formation density so allows the calculation of Poisson's

ratio, which is used to differentiate unconsolidated rocks (Poisson's >0.35) from consolidated rocks (Mari and Porel, 2018). The distribution of velocities and densities can be updated so that an optimum fit between the synthetic and the field data is obtained. This procedure is called full waveform inversion (FWI). The wave equation is used in specific processing procedures, such as tomography and migration.

10.3.2 TIME AVERAGE EQUATION TO ESTIMATE ROCK POROSITY

In porous rocks, the nature of the materials within the pores strongly influences the elastic wave velocity. Water-saturated rocks have different elastic wave velocities compared with gas-saturated rocks.

TABLE 10.1
Seismic Velocities and Densities

Type of Rock or Medium	<i>P</i> Velocity V_P (m/s)	<i>S</i> Velocity V_S (m/s)	Density ρ (gm/cm ³)
Weathered rock	100–300	300–700	1.7–2.4
Dry sand	400–1,200	100–500	1.5–1.7
Wet sands	1,500–4,000	400–1,200	1.9–2.1
Clay	1,100–2,500	200–800	2.0–2.4
Marl/shale	2,000–3,000	750–1,500	2.1–2.6
Sandstone	3,000–4,500	1,200–2,800	2.1–2.4
Limestone	3,500–6,000	2,000–3,300	2.4–2.7
Chalk	2,300–2,600	1,100–1,300	1.8–2.3
Salt	4,500–5,500	2,500–3,100	2.1–2.3
Anhydride	4,000–5,500	2,200–3,100	2.9–3.0
Dolomite	3,500–6,500	1,900–3,600	2.5–2.9
Granite	4,500–6,000	2,500–3,300	2.5–2.7
Basalt	5,000–6,000	2,800–3,400	2.7–3.1
Coal	2,200–2,700	1,000–1,400	1.3–1.8
Water	1,450–1,500	—	1
Ice	3,400–3,800	1,700–1,900	0.9
Oil	1,200–1,250	—	0.6–0.9

Source: After Lavergne (1989).

Seismic velocities can be used to estimate porosity using the time average equation:

$$\frac{1}{v} = \frac{\varphi}{v_f} + \frac{(1-\varphi)}{v_m} \quad (10.4)$$

where v = P wave velocity for a rock, φ = fractional porosity, v_f = acoustic velocity in the pore fluid, and v_m = acoustic velocity in the rock.

10.4 RAY PATHS IN LAYERED MEDIA

At an interface between two rock layers, there is generally a change in propagation velocity resulting from the difference in physical properties of the two layers. At such an interface, the energy within an incident seismic pulse is partitioned into transmitted and reflected pulses. The relative amplitudes of the transmitted and reflected pulses depend on the velocities v , densities ρ , and the angle of incidence. The energy of a seismic wave is proportional to the square of its amplitude.

10.4.1 REFLECTION AND TRANSMISSION OF NORMALLY INCIDENT SEISMIC RAYS

In practice, reflection seismology is carried out at comparatively small angles of incidence. At normal incidence on an interface, a P wave excites no tangential stresses or displacements, and no shear waves are induced. The total energy of the refracted (transmitted) and reflected rays must be equal to the energy of the incident ray. The partitioning of energy between the reflected and refracted P waves then becomes much simpler. It depends on a property of each medium, known as its acoustic impedance, Z , which is defined as the product of the density ρ of the medium and its P wave velocity v . Thus, we have:

$$Z = \rho v = \text{acoustic impedance} \quad (10.5)$$

The smaller the contrast in acoustic impedance across the rock interface, the greater the portion of energy refracted through the interface. The more energy is reflected, the greater the contrast. The solutions of the equations for the amplitudes A_1 and A_2 of the reflected and refracted P waves, respectively, in terms of the amplitude A_0 of the incident wave are given by

$$R = \frac{A_1}{A_0} = \text{reflection coefficient} \quad (10.6)$$

$$\text{or } R = \frac{\rho_2 v_2 - \rho_1 v_1}{\rho_2 v_2 + \rho_1 v_1} = \frac{Z_2 - Z_1}{Z_2 + Z_1} \quad \text{where } -1 < R < 1 \quad (10.7)$$

$$\text{and } T = \frac{A_2}{A_0} = \text{transmission coefficient} \quad (10.8)$$

$$\text{or } T = 2 \frac{Z_1}{Z_2 + Z_1} = \frac{2\rho_1 v_1}{\rho_2 v_2 + \rho_1 v_1} \quad (10.9)$$

As we know, the energy of a wave is proportional to the square of its amplitude. The fraction E_r of the incident energy that is reflected is given by the square of R ; the fraction E_t that is transmitted is equal to $(1 - E_r)$. When the incident wave is reflected at the surface of a medium with higher seismic impedance ($Z_2 > Z_1$), the reflection coefficient R is positive. This means that the reflected wave is in phase with the incident wave. However, if the wave is incident on a medium with lower seismic impedance ($Z_2 < Z_1$), the reflection coefficient will be negative. This implies that the reflected wave is 180° out of phase with the incident wave. The fraction of energy reflected from an interface is equal to R^2 and therefore does not depend on whether the incidence is from the medium of higher or lower seismic impedance.

10.4.2 REFLECTION AND REFRACTION OF THE OBLIQUELY INCIDENT RAYS

In the case of incident waves, reflected and transmitted waves are generated as described in the case of normal incidence. When a P wave is incident at an oblique angle on a plane surface, four types of waves are generated:

- Reflected and transmitted P waves
- Reflected and transmitted S waves

Let the angles between the normal to the interface and the ray paths of the P and S waves in medium 1 be i_p and i_s , respectively, and the corresponding angles in medium 2 be r_p and r_s (Figure 10.3). Applying Snell's law to both the reflected and refracted P and S waves gives

$$\frac{\sin i_p}{\alpha_1} = \frac{\sin i_s}{\beta_1} = \frac{\sin r_p}{\alpha_2} = \frac{\sin r_s}{\beta_2} \Rightarrow \frac{\sin i_p}{\sin r_p} = \frac{\alpha_1}{\alpha_2} \quad (10.10)$$

By similar reasoning, it is evident that an incident SV wave also generates vibrations that have components normal and parallel to the interface and will set up refracted and reflected P and SV waves. The situation is different for an incident SH wave, which has no component of motion normal to the interface. In this case, only refracted and reflected SH waves are created.

10.4.2.1 Subcritical and Supercritical Reflections, and Critical Refraction

Let O be a seismic source near the surface of a uniformly thick horizontal layer with P wave velocity α_1 that lies on top of a layer with higher velocity α_2 (Reynolds, 1998). Consider what happens to seismic rays that leave O and arrive at the boundary with all possible angles of incidence.

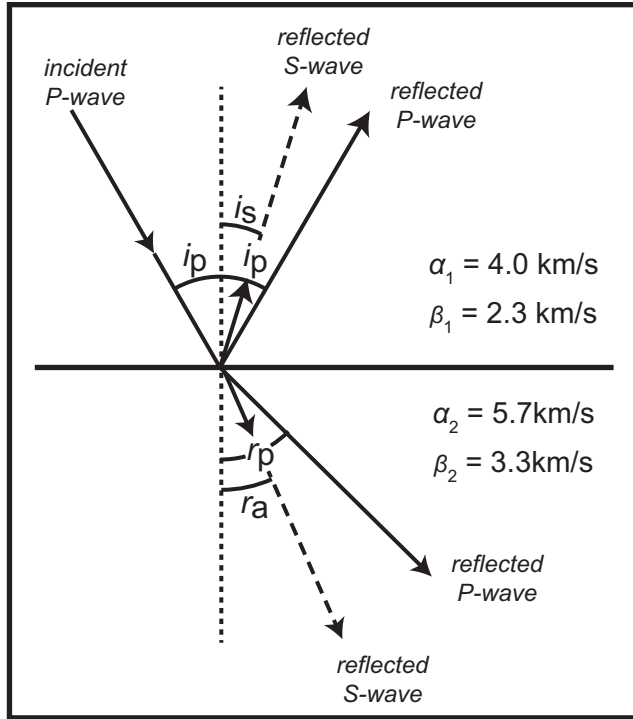


FIGURE 10.3 The generation of reflected and refracted P and S waves from a P wave incident on a plane interface.

Source: Reynolds (1998).

The simplest ray is that which travels vertically to meet the boundary with zero angle of incidence at the point N . This normally incident ray is partially reflected back along its track and partially transmitted vertically into the next medium without change of direction. As the angle of incidence increases, the point of incidence moves from N towards C . The transmitted ray experiences a change of direction according to Snell's law of refraction, and the ray reflected to the surface is termed a subcritical reflection.

The ray that is incident on the boundary at C is called the critical ray because it experiences critical refraction. It encounters the boundary with a critical angle of incidence. The corresponding refracted ray makes an angle of refraction of 90° with the normal to the boundary. As a result, it travels parallel to the boundary in the top of the lower layer with faster velocity α_2 . The sine of the angle of refraction of the critical ray is unity, and we can calculate the critical angle, i_c , by applying Snell's law:

$$\frac{\sin i_c}{\sin 90^\circ} = \frac{\alpha_1}{\alpha_2} \Rightarrow \sin i_c = \frac{\alpha_1}{\alpha_2} \quad (10.11)$$

The critical ray is accompanied by a critical reflection. It reaches the surface at a critical distance (x_c) from the source at O . The reflections that arrive inside the critical distance are called subcritical reflections. At angles up to the critical angle, refracted rays pass into the lower medium, but for rays incident at angles greater than the critical angle, refraction is no longer possible. The seismic rays that are incident more obliquely than the critical angle are reflected almost completely. These reflections are termed supercritical reflections, or simply, wide-angle reflections. They lose little energy to refraction and are thus capable of travelling large distances from the source in the upper medium. Supercritical reflections are recorded with strong amplitudes on seismograms at distant stations.

10.5 SEISMIC DATA ACQUISITION SYSTEMS

The fundamental purpose of seismic surveys is to accurately record ground motion caused by known sources in a known location. The record of ground motion with time constitutes a seismogram (Kirsch, 2006). The essential instrumental requirements are to:

- Generate a seismic pulse with a suitable source
- Detect the seismic waves in the ground with a suitable transducer
- Record and display seismic wave forms on a suitable seismograph

10.5.1 SEISMIC SOURCES

A *seismic source* is a device that generates controlled seismic energy used to perform both reflection and refraction seismic surveys. A seismic source can be simple, such as a dynamite, or it can use more sophisticated technology, such as a specialized airgun. Seismic sources can provide single pulses or continuous sweeps of energy, generating seismic waves, which travel through a medium such as water or layers of rocks. Some of the waves then reflect and refract and are recorded by receivers, such as geophones or hydrophones.

Seismic sources may be used to investigate shallow subsoil structure, for engineering site characterization, or to study deeper structures, either in the search for petroleum and mineral deposits or to map subsurface faults or for other scientific investigations. The returning signals from the sources are detected by seismic sensors (geophones or hydrophones) in known locations relative to the position of the source. The recorded signals are then subjected to specialist processing and interpretation to yield comprehensible information about the subsurface. A seismic source signal has the following characteristics:

- Generates an impulse signal
- Band-limited
- Generated waves are time-varying

The generalized equation that shows all these properties is

$$s(t) = \beta e^{-\alpha t^2} \sin(2\pi f_{\max} t) \quad (10.12)$$

where $f_{\max}$ is the maximum frequency component of the generated waveform.

10.5.1.1 Types of Sources

10.5.1.1.1 Hammer

The most basic seismic source is a sledgehammer, either striking the ground directly or, more commonly, striking a metal plate on the ground, known as hammer-and-plate. Useful for seismic refraction surveys down to about 20 m below the surface.

10.5.1.1.2 Drop-Weight/Accelerated Weight

In 1953, the weight-dropping thumper technique was introduced as an alternative to dynamite sources. A thumper truck (or weight-drop) truck is a vehicle-mounted ground impact system which can be used to provide a seismic source. A heavy weight is raised by a hoist at the back of the truck and dropped, generally about 3 m, to impact (or ‘thump’) the ground (Sheriff, 2002a, 2002b). To augment the signal, the weight may be dropped more than once at the same spot; the signal may also be increased by thumping at several nearby places in an array whose dimensions may be chosen to enhance the seismic signal by spatial filtering.

More advanced thumpers use a technology called ‘accelerated weight drop’ (AWD), where a high-pressure gas (min. 7 MPa, or 1,000 lbf/in²) is used to accelerate a heavyweight hammer (5,000 kg, or 11,000 lb) to hit a base plate coupled to the ground from a distance of 2 to 3 m (6 ft 7 in to 9 ft 10 in). Several thumps are stacked to enhance signal-to-noise ratio. AWD allows both more energy and more control of the source than gravitational weight-drop, providing better depth penetration and control of signal frequency content. Thumping may be less damaging to the environment than firing explosives in shot holes (Chinenyeze, 2015), though a heavily thumped seismic line with transverse ridges every few metres might create a long-lasting disturbance of the soil. An advantage of the thumper (later shared with vibroseis), especially in politically unstable areas, is that no explosives are required.

10.5.1.1.3 Explosives

Explosives most widely used as seismic sources are known as gelatine dynamites (‘Dynamite’, Photo Wiki; E-mail: info@gangalib.org; www.gangalib.org>Alfred-Noble). These dynamites are placed into three subcategories: straight gelatines, in which nitroglycerine, also known as glyceryl trinitrate, with the chemical formula $C_3H_5(ONO_2)_3$, is the active component; ammonia gelatines, in which ammonia nitrite, with chemical formula NH_4NO_3 , is the active component; and semigelatines, in which the composition consists mostly of nitroglycerine (Stark, 2010).

Upon detonation, explosives release large volumes of expanding gas very quickly (‘Explosive chemical product’, *Encyclopaedia Britannica*), forcing great pressure to the surroundings in the form of seismic waves (Madu et al., 2017). Using explosives as seismic sources has been in practice for decades because of the reliability and energy efficiency they provide (Yordkayhun et al., 2009). Such sources are most commonly used on land and swampy environments because of high thickness in sediments (Madu et al., 2017). Typical charge sizes used in the field for reflection surveys are 0.25 kg to 100 kg for single-hole sources, 0.25 kg to 250 kg or more for multiple-hole sources, and may reach 2,500 kg or more for refraction surveys (Stark, 2010).

Though dynamites and other explosives are efficient seismic sources because of their reduced costs, ease of transport in difficult terrains, and lack of regular maintenance compared to other sources (Strobbia et al., 2009), the use of explosives is becoming restricted in certain areas, causing decline and increasing popularity for alternative seismic sources (Yordkayhun et al., 2009). For instance, hexanitrostilbene (HNS) was the main explosive fill in the thumper mortar round canisters used as part of the Apollo lunar active seismic experiments (James et al., 1979) (Wikimedia Commons, file: AP16 pse.jpg). Generally, the explosive charges are placed between 6 and 76 m (20 and 250 ft) belowground, in a hole that is drilled with dedicated drilling equipment for this purpose. This type of seismic drilling is often referred to as ‘shot hole drilling’. A common drill rig used for ‘shot hole drilling’ is the ARDCO C-1000 drill mounted on an ARDCO K 4 × 4 buggy. These drill rigs often use water or air to assist the drilling.

10.5.1.1.4 Airgun

Sound has been used as a tool for imaging geological structures on land and in water for more than 50 yrs. Compressed air sources, referred to as *airguns* (the amplitude of the sound increases in proportion to the cube root of the volume of the airgun), have been the dominant marine sound source since the 1960s (Parkes and Hatton, 1986). A typical compressed air source (‘airgun’) consists of two air chambers surrounding a piston/shuttle (Gisiner, 2016). When the pressure is equal in the two chambers, the ports are blocked by the piston. When the air from one chamber is redirected via a solenoid-activated alternative pathway, the piston is pushed out of the way, allowing the air to escape. The escaping air coalesces into a bubble, thereby generating sound by the ensuing expansions and contractions of the released bubble. Whether on land or in the water, the basic principle is that the acoustic energy from the sound source is reflected and refracted by the rock layers beneath the surface back to the receivers, thereby enabling geophysicists to reconstruct an ‘image’ of the underlying geology, in a way that is analogous to medical ultrasonic imaging. On land, the acoustic

energy comes from buried explosives or vibratory sources that are in contact with the ground; returned vibrations are received by geophones (Sheriff and Geldart, 1995). Little or no energy is transmitted to the air to be perceived as sound.

10.5.1.1.5 Seismic Vibrator

A *seismic vibrator* propagates energy signals into the Earth over an extended period of time as opposed to the near-instantaneous energy provided by impulsive sources. It is one of a number of seismic sources used in reflection seismology (Sheriff, 2002a, 2002b). The data recorded in this way must be correlated to convert the extended source signal into an impulse. The source signal using this method was originally generated by a servo-controlled hydraulic vibrator or shaker unit mounted on a mobile base unit, but electromechanical versions have also been developed (NiBZH, own work: file: nomad 90 vibrating.jpg., from Wikimedia Commons).

10.5.2 TRANSDUCERS

The conversion of the ground motion to an electrical signal requires a transducer which is sensitive to ground motion. Devices used on land to detect seismic ground motions are known as seismometers or geophones (image provided by Sensor Nederland: www.geophone.com, hydrophone in waterborne surveys). The relative movement of the magnet with respect to the coil results in a small voltage. Typically, the amplification of the ground is many orders of magnitude but accomplished on a relative basis. The absolute value of particle acceleration cannot be determined unless the geophones are calibrated. Geophones are chosen for their frequency band response. The line, spread, or string of phones may contain one to scores of sensors, depending on the type of survey. The individual channel of recording normally will have a single phone. Multiple phones per channel may aid in reducing wind noise or air blast or in amplifying deep reflections.

10.5.3 SEISMOGRAPH/SEISMOMETER

The equipment that records input geophone voltages in a timed sequence is the *seismograph*. A record produced by a seismograph on a display screen or paper printout is called a *seismogram* ('a seismogram being recorded by a seismograph at Weston observatory in Massachusetts', Wikipedia). Current practice uses seismographs that store the channels' signals as digital data at a discrete time. Earlier seismographs would record directly to paper or photographic film. Stacking, inputting, and processing the vast volumes of data and archiving the information for the client virtually require digital seismographs. The seismograph system may be an elaborate amalgam of equipment to trigger or sense the source, digitize geophone signals, store multichannel data, and provide some level of processing display. A sophisticated seismograph equipment is not normally required for engineering and environmental surveys. One major exception is the equipment for subbottom surveys or nondestructive testing of pavements.

An important criterion of the seismograph is the dynamic range, which is defined as the ratio of the largest to the smallest recoverable signal. Signal intensity is usually expressed in decibel (dB) units, which is the logarithm of amplitude or power ratio, defined as $20\log_{10}$ of the amplitude ratio or $10\log_{10}$ of the power ratio. Seismologists usually use amplitude rather than power. An increase of 6 dB in signal approximately doubles the amplitude; 60 dB in signal is a factor of 1,000 in amplitude, and 120 dB is a factor of a million. Today, seismographs with dynamic ranges of 100 dB or more are available, and very small signals may be enhanced by digital signal processing. With seismographs of dynamic range less than 100 dB, Knapp and Steeples (1986b) recommend a balancing of the spectrum of the data during recording to detect weak but high-frequency signals.

10.6 SEISMIC REFLECTION METHOD

10.6.1 INTRODUCTION AND GENERAL CONSIDERATIONS

Seismic reflection is the most widely used geophysical technique. Important details about the geometry of the structure and about the physical properties can be derived. Its predominant applications are hydrocarbon exploration and research into crustal structure with depths of penetration of several kilometres. Since 1980, the method has been applied increasingly to engineering and environmental investigations (<200 m) for mapping quaternary deposits, buried valleys, and shallow faults; hydrogeological studies of aquifers; shallow coal exploration; and preconstruction ground investigations for pipe, cables, and sewerage schemes. The essence of the seismic reflection technique is to measure the time taken for a seismic wave to travel from a source down into the ground, where it is reflected back to the surface and then detected at a receiver. The time is known as the two-way travel time (TWTT). The most important problem in seismic reflection surveying is the translation of TWTT to depth. While travel times are measured, the one parameter that most affects the conversation of depth is seismic velocity (two unknown depth and velocity). The velocity is given by the following formula:

$$v_i = \frac{Z_i}{\tau_i} \quad (10.13)$$

where v_i is the interval velocity, Z_i the thickness of the interval, and τ_i the one-way travel time. The travel path of a reflected wave is given in Figure 10.4 (Kearey and Brooks, 2002).

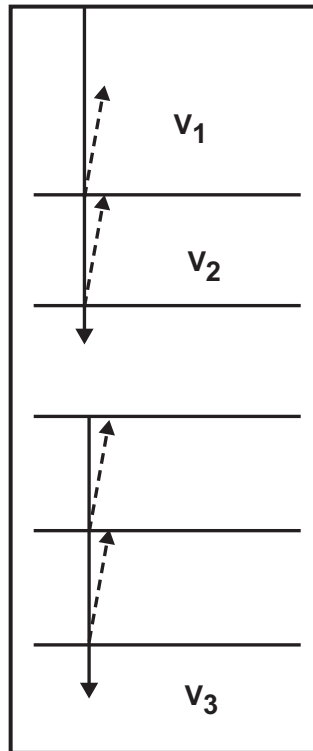


FIGURE 10.4 Vertical reflected ray paths in a horizontally layered ground.

The interval velocity may be averaged over several depth intervals to yield an average velocity $\bar{v}$. We have

$$\bar{v} = \frac{\sum_{i=1}^n Z_i}{\sum_{i=1}^n \tau_i} = \frac{\sum_{i=1}^n v_i \tau_i}{\sum_{i=1}^n \tau_i} \Rightarrow \bar{v} = \frac{Z_n}{T_n} \quad (10.14)$$

where Z_n is the total thickness of the top n layers, and T_n is the total one-way travel time through the n layers.

10.6.1.1 Reflection at a Single Horizontal Reflector

The simplest case of seismic reflection is the two-dimensional reflection at a horizontal boundary (Figure 10.5). Let the reflecting bed be at depth d below the shot point S (Kearey and Brooks, 2002). The ray that strikes the boundary at R is reflected to the surface and recorded by a geophone at the point G so that the angles of incidence and reflection are equal. Let G be at a horizontal distance x from the shot point. If the P wave velocity is V , the first signal received at G is from the direct wave that travels directly along SG . Its travel time is given by $t_d = \frac{x}{V}$. It is important to keep in mind that the direct wave is not a surface wave but a body wave that travels parallel to and just below the surface of the top layer. The travel time t of the reflected ray SRG is $\frac{SR + RG}{V}$. However, SR and RG are equal, and therefore,

$$t = \frac{2}{V} \sqrt{d^2 + \frac{x^2}{4}} \quad (10.15)$$

$$\text{or } t = \frac{2d}{V} \sqrt{1 + \frac{x^2}{4d^2}} = t_0 \sqrt{1 + \frac{x^2}{4d^2}} \quad (10.16)$$

At $x = 0$, the travel time corresponds to the vertical echo from the reflector; this ‘echo time’ is given by t_0 . The quantity under the square root in eq. 10.16 determines the curvature of the t - x curve and is called the normal moveout factor. It arises because the ray reaching a geophone at a horizontal distance x from the shot point has not travelled vertically between it and the reflector. Squaring both sides of eq. 10.16 and rearranging the terms give

$$\frac{t^2}{t_0^2} - \frac{x^2}{4d^2} = 1 \quad (10.17)$$

This is the equation of a hyperbola (Figure 10.23) that is symmetrical about the vertical time axis, which it intersects at t_0 . For large distances from the shot point ($x \gg 2d$), the travel time of the reflected ray approaches the travel time of the direct ray, and the hyperbola is asymptotic to the two lines $t = \pm \frac{x}{V}$. A principal goal of seismic reflection profiling is usually to find the vertical distance (d) to a reflecting interface. This can be determined from t_0 , the two-way reflection travel time recorded by a geophone at the shot point, once the velocity V is known. One way of determining the velocity is by comparing t_0 with the travel time t_x to a geophone at distance x . In reflection seismology, the geophones are laid out close to the shot point, and the assumption is made that the geophone distance is much less than the depth of the reflector ($x \ll d$). Eq. 10.16 becomes

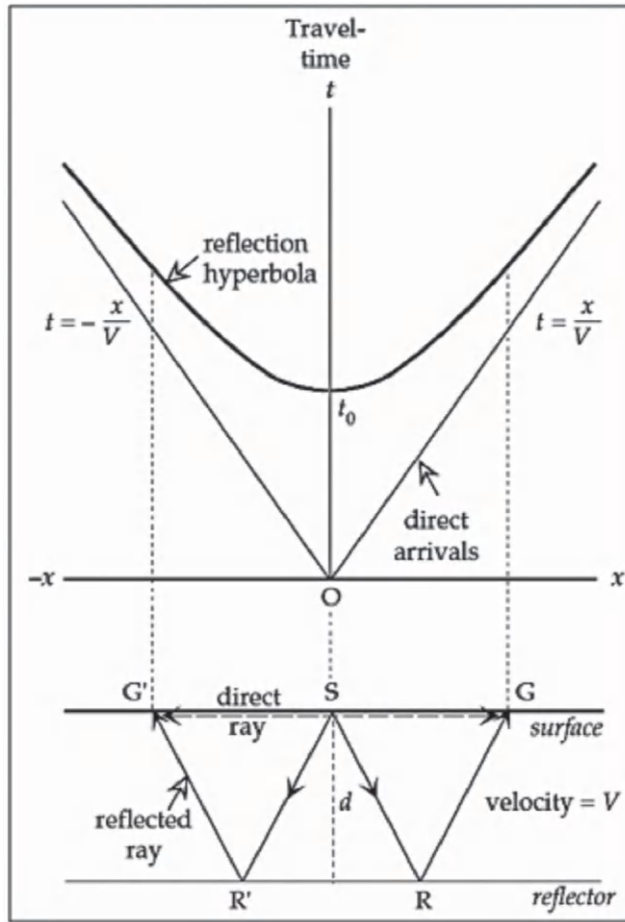


FIGURE 10.5 The 'travel time vs distance' curve for reflections from a horizontal boundary is a hyperbola. The vertical reflection time t_0 is the intercept of the hyperbola with the travel time axis.

$$t_x = t_0 \left\{ 1 + \left(\frac{x}{2d} \right)^2 \right\}^{\frac{1}{2}} = t_0 \left\{ 1 + \frac{1}{2} \left(\frac{x}{2d} \right)^2 + \dots \right\} = t_0 \left\{ 1 + \frac{1}{2} \left(\frac{x}{Vt_0} \right)^2 \right\} \quad (10.18)$$

The difference between the travel time t_x and the shot point travel time t_0 is the normal move-out, $\Delta t_n = t_x - t_0$. By rearranging eq. 10.18, we get

$$\Delta t_n = \frac{x^2}{2V^2 t_0} \quad (10.19)$$

The echo time t_0 and the normal moveout time Δt_n are found from the reflection data. The distance x of the geophone from the shot point is known, and therefore, the layer velocity V can be determined. The depth d of the reflecting horizon can then be found by using the formula for the echo time.

An alternative way of interpreting reflection arrival times becomes evident when eq. 10.17 is rearranged in the form

$$t^2 = t_0^2 + \frac{x^2}{V^2} \quad (10.20)$$

A plot of t^2 against x^2 is a straight line that has slope $\frac{1}{V^2}$. Its intercept with the t^2 axis gives the square of the echo time, t_0 , from which the depth d to the reflector can be found once the velocity V is known. The record at each geophone will contain reflections from several reflectors. For the first reflector, the velocity determined by the $t^2 - x^2$ method is the true interval velocity of the uppermost layer, V_1 , which, in conjunction with t_{01} , the first echo time, gives the thickness d_1 of the top layer. However, the ray reflected from the second interface has travelled through the first layer with interval velocity V_1 and the second layer with interval velocity V_2 . The velocity interpreted in the $t^2 - x^2$ method for this reflection, and for reflections from all deeper interfaces, is an average velocity. If the incident and reflected rays travel nearly vertically, the average velocity $V_{a,n}$ for the reflection from the n^{th} reflector is given by

$$V_{a,n} = \frac{d_1 + d_2 + d_3 + \dots + d_n}{t_1 + t_2 + t_3 + \dots + t_n} = \frac{\sum_{i=1}^n d_i}{\sum_{i=1}^n t_i} \quad (10.21)$$

where d_i is the thickness and t_i the interval travel time for the i^{th} layer.

The $t^2 - x^2$ method is a simple way of estimating layer thicknesses and average velocities for a multilayered Earth (Figure 10.6). The slope of the second straight line gives $V_{a,2}$, which is used with the appropriate echo time t_{02} to find the combined depth D_2 to the second interface, given by $D_2 = d_1 + d_2 = V_{a,2}t_{02}$; d_1 is known, and so d_2 can be calculated. The two-way travel time in the second layer is $(t_{02} - t_{01})$, and thus the interval velocity V_2 can be found. In this way, the thicknesses and interval velocities of the deeper layers can be determined successively. In fact, of course, the rays do not travel vertically but are bent as they pass from one layer to another (Figure 10.6). Moreover, the elastic properties of a layer are rarely homogeneous, so that the seismic velocity is variable and the ray path in the layer is curved (Kearey and Brooks, 2002). Exploration seismologists have found it possible to compensate for these effects by replacing the average velocity with the root mean square velocity V_{rms} , defined by

$$V_{\text{rms}}^2 = \frac{\sum_{i=1}^n V_i^2 t_i}{\sum_{i=1}^n t_i} \quad (10.22)$$

where V_i is the interval velocity and t_i the travel time for the i^{th} layer.

10.6.1.2 Single Dipping Reflector

When the reflecting interface is inclined at an angle to the horizontal, as in Figure 10.7, the shortest distance d between the shot points and the reflector is the perpendicular distance to the inclined plane. The paths of reflected rays on the down-dip side of the shot point are longer than those on the up-dip side; this has corresponding effects on the travel times. The rays obey the laws of reflection optics and appear to return to the surface from the point S' , which is the image point of the shot point with respect to the reflector. The travel time t through the layer with velocity V is readily found with the aid of the image point. For example, for the ray SRG , recorded by a geophone on the surface at G , we get

$$t = \frac{SR + RG}{V} = \frac{S'R + RG}{V} = \frac{S'G}{V} \quad (10.23)$$

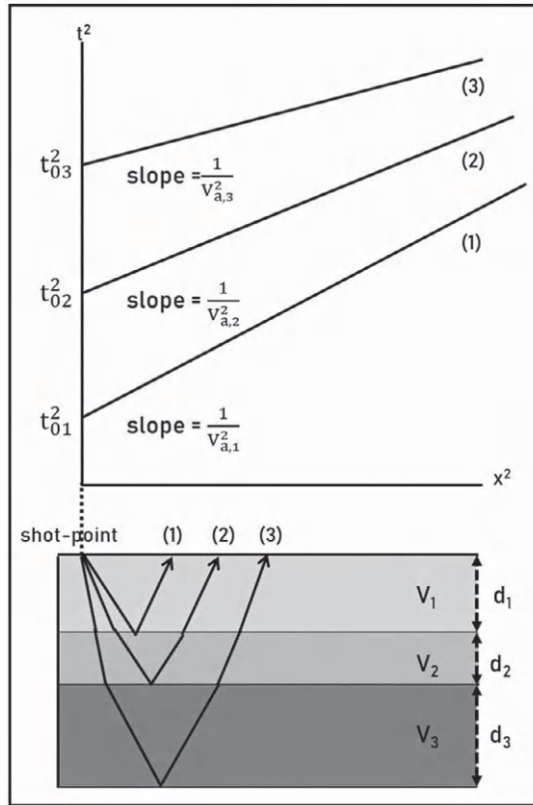


FIGURE 10.6 Illustration of a ' $t^2 - x^2$ ' plot for near-vertical reflections from three horizontal reflectors; $V_{a,2}$ is the true velocity V_1 of layer 1, but $V_{a,2}$ and $V_{a,3}$ are 'average' velocities that depend on the true velocities and the layer thicknesses.

The image point S' is as far behind the reflector as the shot point is in front: $S'S = 2d$. In triangle $S'SG$, the side SG equals the geophone distance x and the obtuse angle $S'SG$ equals $(90^\circ + \theta)$. If we apply the law of cosines to the triangle $S'SG$, we can solve for $S'G$ and substitute the answer in eq. 10.23. This gives the travel time of the reflection from the inclined boundary:

$$t = \frac{1}{V} \sqrt{x^2 + 4xd\sin\theta + d^2} \quad (10.24)$$

Eq. 10.24 is the equation of a hyperbola whose axis of symmetry is vertical, parallel to the t axis. For a flat reflector, the hyperbola is symmetric about the t axis (Figure 10.5) and the minimum travel time (echo time) corresponds to the vertical reflection below the shot point ($x = 0$). For an inclined reflector the minimum travel time t_m is no longer the perpendicular path to the reflector, which would give the travel time t_0 in Figure 10.7. Although the perpendicular path is the shortest distance from shot point to reflector, it is not the shortest path of a reflected ray between the shot point and a geophone (Kearey and Brooks, 2002). The shortest travel time is recorded by the geophone at a horizontal distance x_m on the up-dip side of the shot point. The coordinates (x_m, t_m) of the minimum point of the travel time hyperbola are

$$x_m = -2d\sin\theta \text{ and } t_m = \frac{2d\cos\theta}{V} \quad (10.25)$$

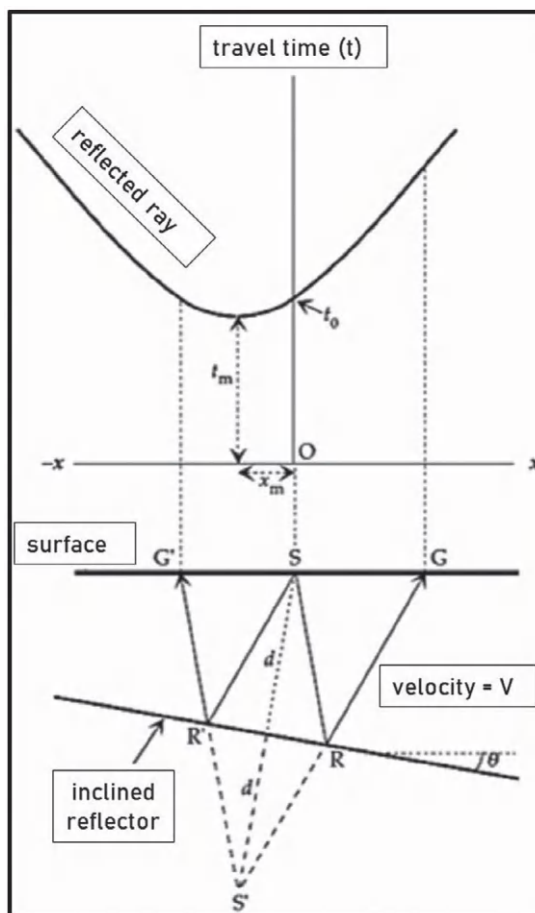


FIGURE 10.7 The 'travel time vs distance' curve for an inclined reflector is also a hyperbola with vertical axis (see Figure 10.23), but the minimum travel time (t_m) is measured at distance X_m from the shot point.

In practice, it is not known a priori whether a reflector is horizontal or inclined. If reflection records are not corrected for the effect of layer dip, an error results in plotting the positions of dipping beds. The shot point travel time t_0 gives the direct distance to a reflector, but the path along which the echo has travelled is not known. Consider the geometry of the inclined boundary in Figure 10.8. First arrival reflections recorded for shot points P , Q , and R come from the true reflection points A , B , and C . If the computed reflector depths are plotted directly below the shot points at A' , B' , and C' , the dipping boundary will appear to lie at a shallower depth than its true position, and the apparent dip of the reflector will be less steep than the true dip (Kearey and Brooks, 2002). This leads to a distorted picture of the underground structure. For example, an anticline appears broader and less steep-sided than it is. Similarly, if the limbs of a syncline dip steeply enough, the first arrivals from the dipping limbs can conceal the true structure (Figure 10.9).

This happens when the radius of curvature of the bottom of the syncline is less than the subsurface depth of its axis. Over the axis of the syncline, rays reflected from the dipping flanks may be the first to reach the shot point geophone. The bottom of the syncline is seen as an upwardly convex reflection between two cusps (Figure 10.10). On an uncorrected reflection record, the appearance of a tight syncline resembles diffraction.

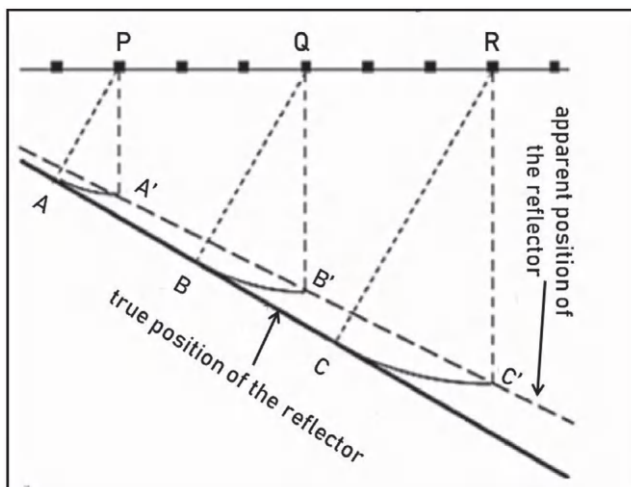


FIGURE 10.8 When the reflector is inclined and depths are plotted vertically under geophone positions, the true reflecting points A, B, and C are mapped at A', B', and C', falsifying the position of the reflector.

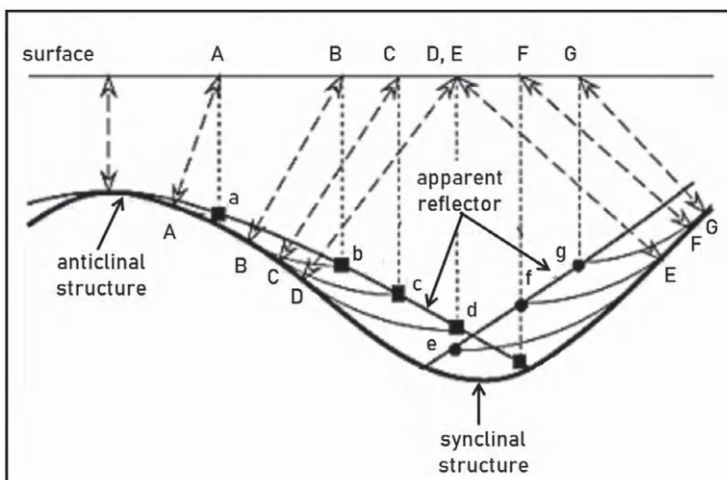


FIGURE 10.9 Paths of reflected rays over an anticline and syncline, showing the false apparent depth to the reflecting surface. True reflection points A–G are wrongly mapped at locations a–g beneath the corresponding shot points.

Source: Lowrie (2007).

Reflection seismic records must be corrected for nonvertical reflections. The correctional process is called migration. It is an essential part of a reflection seismic study. When the reflection events on seismic cross sections are plotted vertically below control points on the surface (e.g. as the two-way vertical travel time to a reflector below the shot point), the section is said to be unmigrated. As discussed earlier, an unmigrated section misrepresents the depth and dip of inclined reflectors. A migrated section is one which has been corrected for nonvertical reflections. It gives a truer picture of the positions of subsurface reflectors. The process of migration is complex and requires prior knowledge of the seismic velocity distribution, which, in an unexplored or tectonically complicated region, is often inadequately known.

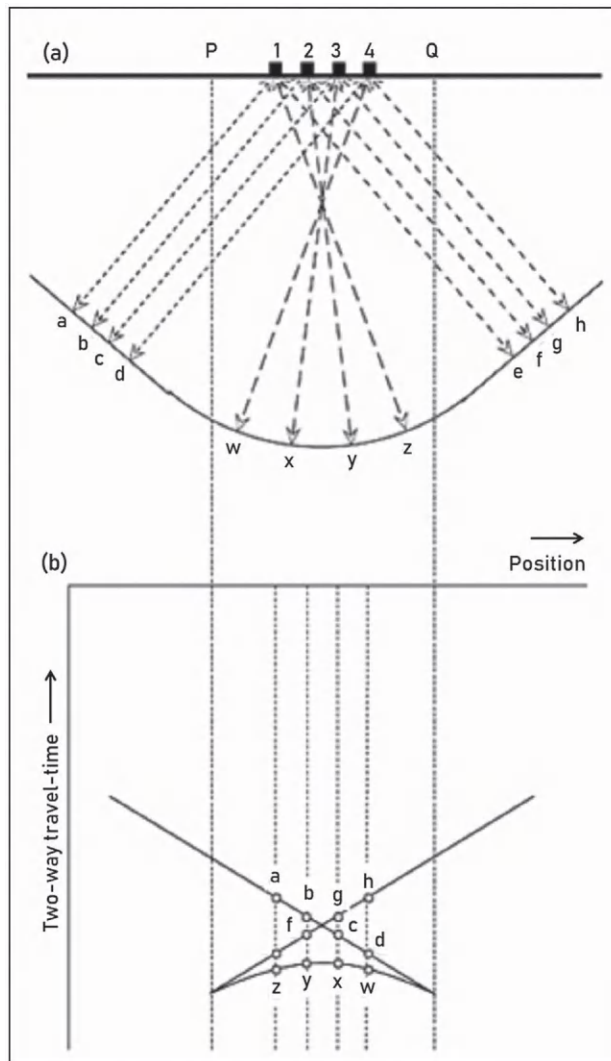


FIGURE 10.10 (a) Paths of rays reflected from both flanks and the trough of a tightly curved syncline. (b) Appearance of the corresponding reflection record; the letters on the cusped feature refer to the reflection points in (a).

Source: Lowrie (2007).

10.6.2 SEISMIC REFLECTION FIELD SURVEY

Several field procedures are in common use. They are distinguished by different layouts of the geophones relative to the shot point. The most routine application of reflection seismology is in continuous profiling, in which the geophones are laid out at discrete distances along a profile through the shot point. To reduce seismic noise, each recording point is represented by a group of interconnected geophones. After each shot, the geophone layout and shot point are moved a predetermined distance along the profile, and the procedure is repeated. Broadly speaking, there are two main variations of this method, depending on whether each reflection point on the reflector is sampled only once (conventional coverage) or more than once (redundant coverage).

The most common form of conventional coverage is a split-spread method (Figure 10.11), in which the geophones are spread symmetrically on either side of the shot point. If the reflector

is flat-lying, the point of reflection of a ray recorded at any geophone is below the point midway between the shot point and the geophone. For a shot point at Q , the rays QAP and QBR that are reflected to geophones at P and R represent extreme cases. The two-way travel time of the ray QAP gives the depth of the reflection point A , which is plotted below the midpoint of QP . Similarly, B is plotted below the midpoint of QR . The split-spread layout around the shot point Q gives the depths of reflection points along AB , which is half the length of the geophone-spread PR . The shot point is now moved to the point R , and the geophones between P and Q are moved to cover the segment RS . From the new shot point R the positions of reflection points in the segment BC of the reflector are obtained. The ray RBQ from shot point R to the geophone at Q has the same path as the ray QBR from shot point Q to the geophone at R . By successively moving the shot point and half of the split-spread geophone layout, a continuous coverage of the subsurface reflector is obtained.

Redundant coverage is illustrated by the common midpoint method, which is routinely employed as a means of reducing noise and enhancing the signal-to-noise ratio. Commonly, 24 to 96 groups of geophones feed recorded signals into a multichannel recorder. The principle of common midpoint coverage is illustrated for a small number of 11 geophone groups in Figure 10.12. When a shot is fired at A ,

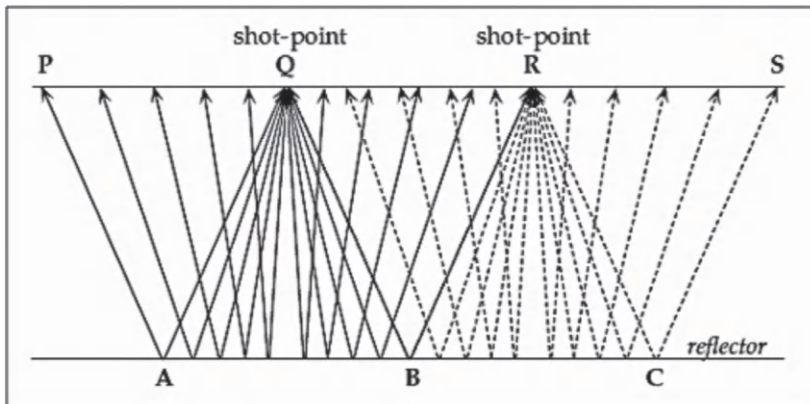


FIGURE 10.11 The split-spread method of obtaining continuous subsurface coverage of a seismic reflector.

Source: Seismic Reflection Surveying. www.geophysik.uni-muenchen.de->studium->Vorlesungen.

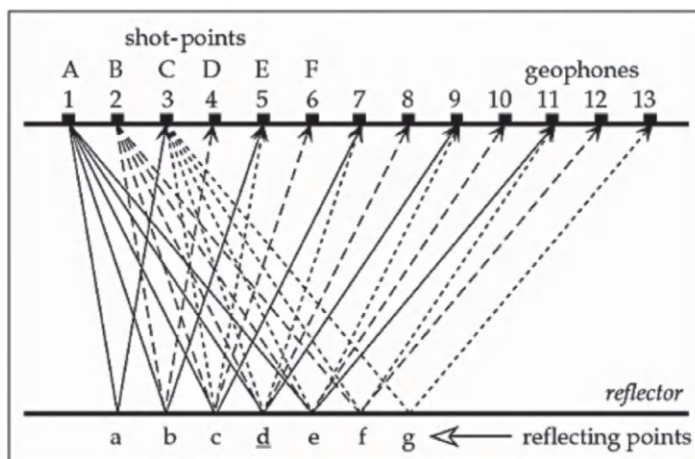


FIGURE 10.12 Common midpoint method of seismic reflection shooting, showing rays from successive shot points at A , B , and C and the repeated sampling of the same point on the reflector (e.g. d) by rays from each shot point.

the signals received at geophones 3–11 give subsurface coverage of the reflector between points a and e . The shot point is now moved to B , which coincides with the position occupied by geophone 2 for the first shot, and the geophone array is moved forward correspondingly along the direction of the profile to positions 4–12. From shot point B , the subsurface coverage of the reflector is between points b and f . The reflector points b to e are common to both sets of data. By repeatedly moving the shot point and geophone array in the described manner, each reflecting point of the interface is sampled multiply. For example, in Figure 10.12, the reflecting point d is sampled multiply by the rays $Ad9$, $Bd8$, $Cd7$, etc. The lengths of these ray paths are different. During subsequent data processing, the reflection travel times are corrected for normal moveout, which is a geometrical effect related to geophone distance from the shot point. The records are then stacked, which is a procedure for enhancing the signal-to-noise ratio.

The fold of a CMP gather is determined by $m = \frac{N}{2n}$, where N is the number of geophones, and n is the number of array spacings the source is moved between shots. The improvement of the $\frac{S}{N}$ ratio from stacking m traces is $\sqrt{m}$, for example, $N = 96$, $n = 8 \Rightarrow \frac{96}{16} = 6$ -fold (Figure 10.13). Improvement of $\frac{S}{N}$ ratio is $\sqrt{6} \approx 2.5$.

The CMP gather lies in the heart of seismic processing for two main reasons. One is the variation of travel time with offset; the moveout will depend only on the velocity of the subsurface

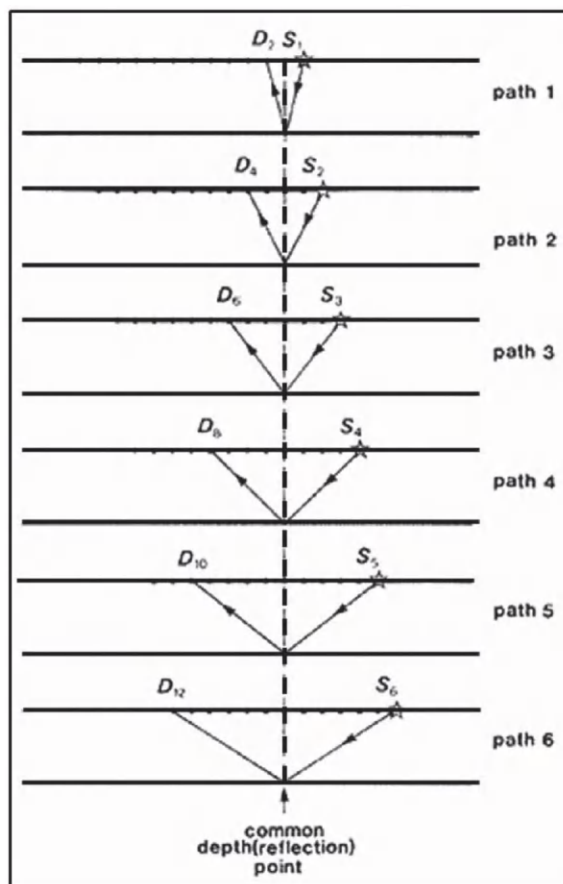


FIGURE 10.13 Sketch rays for the common midpoint (CMP) method.

Source: Telford et al. (1990).

layers (horizontal uniform layers). The subsurface velocity can be derived. Another is, the reflected seismic energy is usually very weak. It is imperative to increase the signal–noise ratio of most data.

10.6.2.1 Vertical and Horizontal Resolution (Fresnel Zone)

Vertical resolution of a seismic pulse is $\left(\frac{1}{4} \text{ to } \frac{1}{8}\right)$ of a wavelength, for example, if $v = 2 \text{ km/s}$ and frequency = 50 Hz, then the resolution is 10 m. Sharpening of the pulse desirable $\rightarrow$ deconvolution $\rightarrow$ towards impulse response of the medium. Horizontal resolution is determined by the detector spacing and the Fresnel zone, where w is $(2z\lambda)^{\frac{1}{2}}$ for $z \gg \lambda$ (Figures 10.14a and b), z the reflection depth (Telford et al., 1991).

10.7 SEISMIC REFRACTION SURVEYING

The method of seismic refraction can be understood by applying the Huygens principle to the critical refraction at the interface between two layers. The seismic disturbance travels immediately

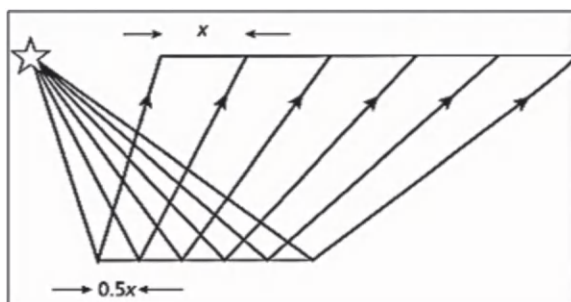


FIGURE 10.14A Ray path of seismic reflection survey.

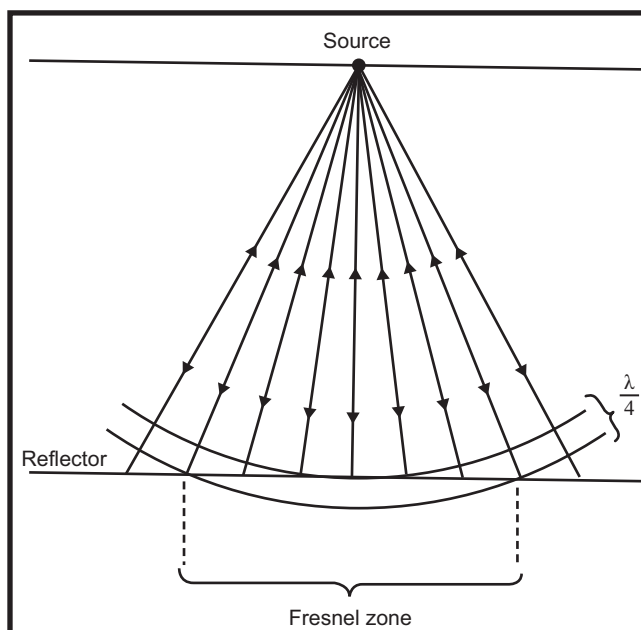


FIGURE 10.14B Seismic reflection survey to show Fresnel zone.

below the interface with the higher velocity of the lower medium. It is called a head wave (or Mintrop wave, after the German seismologist who patented its use in seismic exploration in 1919). The upper and lower media are in contact at the interface, and so the upper medium is forced to move in phase with the lower medium. The vibration excited at the boundary by the passage of the head wave acts as a moving source of secondary waves in the upper layer. The secondary waves interfere constructively (in the same way as a reflected wave is formed) to build plane wave fronts; the ray paths return to the surface at the critical angle within the region of supercritical reflections (Figure 10.15). The doubly refracted waves are especially important for the information they reveal about the layered structure of the deep interior of the Earth (Reynolds, 1998).

10.7.1 REFRACTION AT A HORIZONTAL INTERFACE

The method of refraction seismology is illustrated for the case of the flat interface between two horizontal layers in Figure 10.15. Let the depth to the interface be d and the seismic velocities of the upper and lower layers be V_1 and V_2 , respectively ($V_1 < V_2$). The direct ray from the shot point S is recorded by a geophone G at distance x on the surface after time $\frac{x}{V_1}$. The travel time curve

for the direct ray is a straight line through the origin with slope $m_1 = \frac{1}{V_1}$. The hyperbolic t - x curve

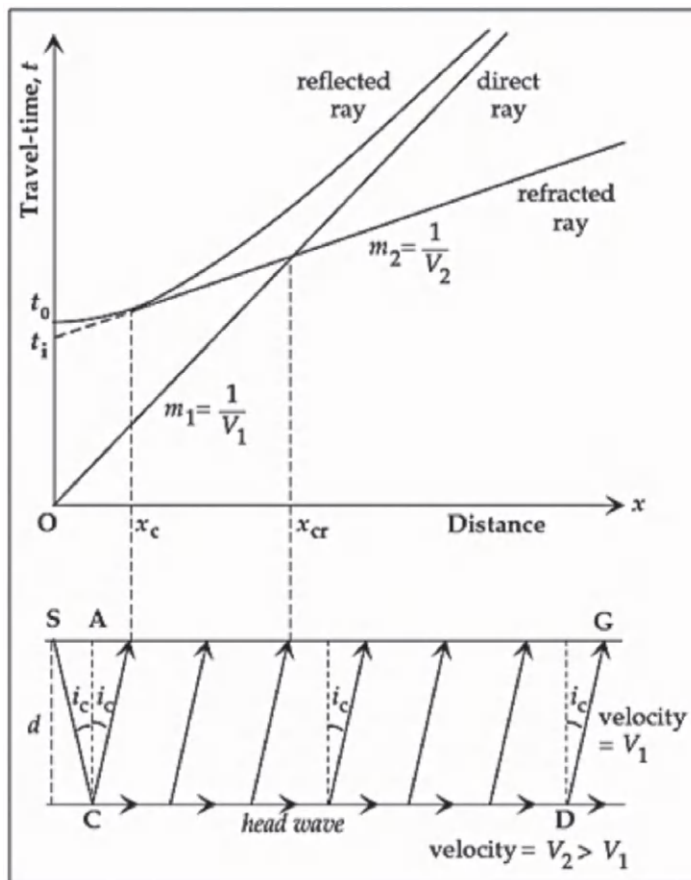


FIGURE 10.15 Travel time vs distance curves for the direct ray and the reflected and refracted rays at a horizontal interface between two layers with seismic velocities V_1 and V_2 .

for the reflected ray intersects the time axis at the two-way vertical reflection ('echo') time t_0 . At great distances from the shot point, the reflection hyperbola is asymptotic to the straight line for the direct ray.

The doubly refracted ray travels along the path SC with the velocity V_1 of the upper layer, impinges with critical angle i_c on the interface at C , passes along the segment CD with velocity V_2 of the lower layer, and returns to the surface along DG with velocity V_1 . The segments SC and DG are equal, $CD = x - 2SA$, and the travel time for the path $SCDG$ can be written as

$$t = \frac{2SC}{V_1} + \frac{CD}{V_2} \quad (10.26)$$

$$\text{That is, } t = \frac{2d}{V_1 \cos i_c} + \frac{x - 2d \tan i_c}{V_2} \quad (10.27)$$

Rearranging the terms and using Snell's law, $\sin i_c = \frac{V_1}{V_2}$, we get for the travel time of the doubly refracted ray

$$t = \frac{x}{V_2} + \frac{2d}{V_1} \cos i_c \quad (10.28)$$

The equation represents a straight line with slope $m_2 = \frac{1}{V_2}$. The doubly refracted rays are only recorded at distances greater than the critical distance x_c . The first arrival recorded at x_c can be regarded as both a doubly refracted ray and a reflection; the travel time line for the head wave is tangential to the reflection hyperbola at x_c . By backward extrapolation, the refraction t - x curve is found to intersect the time axis at the intercept time t_i , given by

$$t_i = \frac{2d}{V_1} \cos i_c = 2d \frac{\sqrt{V_2^2 - V_1^2}}{V_1 V_2} \quad (10.29)$$

Close to the shot point, the direct ray is the first to be recorded. However, the doubly refracted ray travels part of its path at the faster velocity of the lower layer so that it eventually overtakes the direct ray and becomes the first arrival. The straight lines for the direct and doubly refracted rays cross each other at this distance, which is accordingly called the crossover distance, x_{cr} . It is computed by equating the travel times for the direct and refracted rays:

$$\frac{x}{V_1} = \frac{x}{V_2} + 2d \frac{\sqrt{V_2^2 - V_1^2}}{V_1 V_2} \quad (10.30)$$

$$\text{or } x_{cr} = 2d \frac{\sqrt{V_2^2 - V_1^2}}{V_1 V_2} \quad (10.31)$$

Refraction seismology gives the velocities of subsurface layers directly from the reciprocal slopes of the straight lines corresponding to the direct and doubly refracted rays. Once these velocities have been determined, it is possible to compute the depth d to the interface by using either the intercept time t_i or the crossover distance x_{cr} , which can be read directly from the t - x plot:

$$d = \frac{1}{2} t_i \frac{V_1 V_2}{\sqrt{V_2^2 - V_1^2}} \quad (10.32)$$

$$d = \frac{1}{2} x_{cr} \frac{V_1 V_2}{\sqrt{V_2^2 - V_1^2}} \quad (10.33)$$

10.7.2 REFRACTION AT AN INCLINED INTERFACE

In practice, the refracting interface is often not horizontal. The assumption of flat layers then leads to errors in the velocity and depth estimates. When the refractor is suspected to have a dip, the velocities of the beds and the dip of the interface can be obtained by shooting a second complementary profile in the opposite direction. Suppose a refractor dips at an angle θ as in Figure 10.16. Shot points A and B are located at the ends of a geophone layout that covers AB . The ray $ACDB$ from the shot point A strikes the interface at the critical angle i_c at C , runs as a head wave with velocity V_2 along the dipping interface, and the ray emerging at D eventually reaches a geophone at the end of

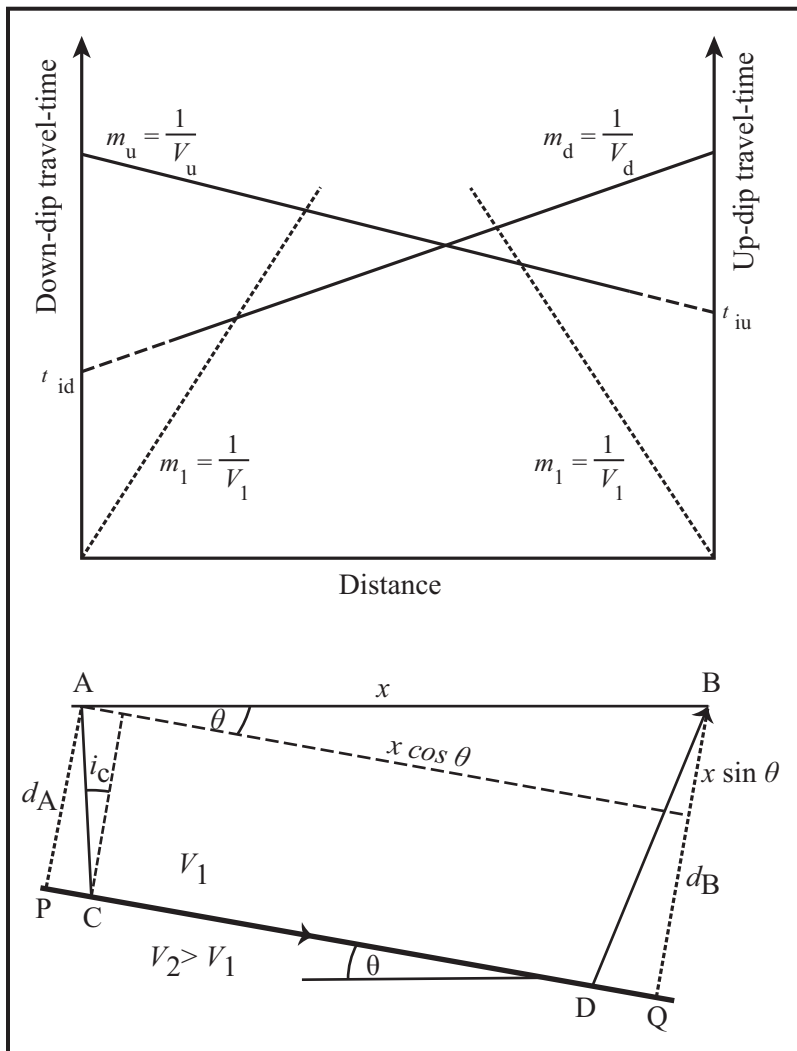


FIGURE 10.16 Travel time vs distance curves of direct and refracted rays for up-dip and down-dip profiles when the refracting boundary dips at angle θ .

the profile at B . During reverse shooting, the ray from the shot point at B to a geophone at A traverses the same path in the reverse direction. However, the t - x curves are different for the up-dip and down-dip shots. Let d_A and d_B be the perpendicular distances from the shot points A and B to the interface at P and Q , respectively. For the downdip shot at A , the travel time to distance x is given by

$$t_d = \frac{AC + DB}{V_1} + \frac{CD}{V_2} \quad (10.34)$$

The geometry of Figure 10.16 gives the following trigonometric relationships:

$$\begin{aligned} \frac{d_A}{AC} &= \cos i_c \Rightarrow AC = \frac{d_A}{\cos i_c}; \quad \frac{d_B}{DB} = \cos i_c \Rightarrow DB = \frac{d_B}{\cos i_c} \\ \frac{PC}{d_A} &= \tan i_c \Rightarrow PC = d_A \tan i_c; \quad \frac{DQ}{d_B} = \tan i_c \Rightarrow DQ = d_B \tan i_c \\ x \cos \theta &= PC + CD + DQ \Rightarrow CD = x \cos \theta - (PC + DQ) \end{aligned} \quad (10.35)$$

$$\text{We have } t_d = \frac{d_A + d_B}{V_1 \cos i_c} + \frac{x \cos \theta - (PC + DQ)}{V_2} \Rightarrow t_d = \frac{d_A + d_B}{V_1 \cos i_c} + \frac{x \cos \theta - (d_A + d_B) \tan i_c}{V_2}$$

$$\text{We have } t_d = \frac{x \cos \theta}{V_2} + \frac{d_A + d_B}{V_1} \cos i_c \quad (10.36)$$

Eq. 10.36 can be simplified by noting that

$$d_B = d_A + x \sin \theta \text{ and } \frac{1}{V_2} = \frac{\sin i_c}{V_1} \quad (10.37)$$

After substitution and gathering the terms, the down-dip travel time is given by

$$t_d = \frac{x \sin i_c \cos \theta}{V_1} + \frac{x \cos i_c \sin \theta}{V_1} + \frac{2d_A \cos i_c}{V_1} = \frac{x}{V_1} \sin(i_c + \theta) + t_{id} \quad (10.38)$$

where t_{id} is the intercept time for the down-dip shot:

$$t_{id} = \frac{2d_A}{V_1} \cos i_c \quad (10.39)$$

The analysis for shooting in the up-dip direction is analogous and gives

$$t_u = \frac{x}{V_1} \sin(i_c - \theta) + t_{iu} \quad (10.40)$$

where t_{iu} is the intercept time for the up-dip shot:

$$t_{iu} = \frac{2d_B}{V_1} \cos i_c \quad (10.41)$$

If the upper layer is homogeneous, the segments for the direct ray will have equal slopes, the reciprocals of which give the velocity V_1 of the upper layer. The segments of the t - x curves

corresponding to the doubly refracted ray are different for up-dip and down-dip shooting. The total travel times in either direction along $ACDB$ must be equal, but the t - x curves have different intercept times. As these are proportional to the perpendicular distances to the refractor below the shot points, the up-dip intercept time t_{iu} is larger than the down-dip intercept time t_{id} . This means that the slope of the up-dip refraction in Figure 10.16 is flatter than the down-dip slope (Reynolds, 1998). If we interpret the reciprocal of the slope as the velocity of the lower medium, we get two apparent velocities, V_d and V_u , given by

$$\frac{1}{V_d} = \frac{1}{V_1} \sin(i_c + \theta) \text{ and } \frac{1}{V_u} = \frac{1}{V_1} \sin(i_c - \theta) \quad (10.42)$$

Once the real velocity V_1 and the apparent velocities V_d and V_u have been determined from the t - x curves, the dip of the interface θ and the critical angle i_c (and from it the true velocity V_2 of the lower layer) can be computed:

$$\theta = \frac{1}{2} \left\{ \sin^{-1} \frac{V_1}{V_d} - \sin^{-1} \frac{V_1}{V_u} \right\} \quad (10.43)$$

$$i_c = \frac{1}{2} \left\{ \sin^{-1} \frac{V_1}{V_d} + \sin^{-1} \frac{V_1}{V_u} \right\} \quad (10.44)$$

If the reciprocal apparent velocities (eq. 10.42) are added, a simple approximation for the true velocity of the lower layer is obtained:

$$\frac{1}{V_d} + \frac{1}{V_u} = \frac{1}{V_1} \{ \sin(i_c + \theta) + \sin(i_c - \theta) \} = \frac{2}{V_1} \sin i_c \cos \theta = \frac{2}{V_2} \cos \theta \quad (10.45)$$

If the refractor dip is small, $\cos \theta \approx 1$ (for example, if $\theta < 15^\circ$, $\cos \theta > 0.96$), and an approximate formula for the true velocity of the second layer is

$$\frac{1}{V_2} \approx \frac{1}{2} \left(\frac{1}{V_d} + \frac{1}{V_u} \right) \quad (10.46)$$

10.8 SEISMIC TOMOGRAPHY

Seismic tomography is a technique for imaging the subsurface of the Earth with seismic waves produced by earthquakes or explosions. P , S , and surface waves can be used for tomographic models of different resolutions based on seismic wavelength, wave source distance, and the seismograph array coverage (Nolet, 1987). The data received at seismometers are used to solve an inverse problem, wherein the locations of reflection and refraction of the wave paths are determined. This solution can be used to create 3-D images of velocity anomalies, which may be interpreted as structural, thermal, or compositional variations. Geoscientists use these images to better understand core, mantle, and plate tectonic processes.

10.8.1 THEORY

Tomography is solved as an inverse problem. Seismic travel time data are compared to an initial Earth model, and the model is modified until the best possible fit between the model predictions and observed data is found. Seismic waves would travel in straight lines if the Earth were of uniform

composition, but the compositional layering, tectonic structure, and thermal variations reflect and refract seismic waves. The location and magnitude of these variations can be calculated by the inversion process, although solutions to tomographic inversions are non-unique. Seismic tomography is similar to medical X-ray computed tomography (CT scan) in that a computer processes receiver data to produce a 3-D image, although CT scans use attenuation instead of travel time difference. Seismic tomography has to deal with the analysis of curved ray paths, which are reflected and refracted within the Earth, and potential uncertainty in the location of the earthquake hypocentre. CT scans use linear X-rays and a known source (seismic tomography using earthquakes to image Earth's interior) (Incorporated Research Institutions for Seismology [IRIS]).

10.8.2 HISTORY

Seismic tomography requires large datasets of seismograms and well-located earthquake or explosion sources. These became more widely available in the 1960s with the expansion of global seismic networks, and in the 1970s when digital seismograph data archives were established. These developments occurred concurrently with advancements in computing power that were required to solve inverse problems and generate theoretical seismograms for model testing (Shearer, 2016).

In 1977, *P* wave delay times were used to create the first seismic array scale 2-D map of seismic velocity. In the same year, *P* wave data were used to determine 150 spherical harmonic coefficients for velocity anomalies in the mantle (Nolet, 1987). The first model using iterative techniques, required when there are a large number of unknowns, was done in 1984. This built upon the first radially anisotropic model of the Earth, which provided the required initial reference frame to compare tomographic models for iteration. Initial models had resolution of $\sim 3,000$ to $5,000$ km, as compared to the few hundred kilometre resolution of current models (Romanowicz, 2003). Seismic tomographic models improve with advancements in computing and expansion of seismic networks. Recent models of global body waves used over 10^7 travel times to model 10^5 to 10^6 unknowns (Rawlinson et al., 2010).

10.8.3 PROCESS

Seismic tomography uses seismic records to create 2-D and 3-D images of subsurface anomalies by solving large inverse problems such that it generates models consistent with observed data. Various methods are used to resolve anomalies in the crust and lithosphere, shallow mantle, whole mantle, and core based on the availability of data and types of seismic waves that penetrate the region at a suitable wavelength for feature resolution. The accuracy of the model is limited by the availability and accuracy of seismic data, wave type utilized, and assumptions made in the model. *P* wave data are used in most local models and global models in areas with sufficient earthquake and seismograph density. *S* and surface-wave data are used in global models when this coverage is not sufficient, such as in ocean basins and away from subduction zones. First arrival times are the most widely used, but models utilizing reflected and refracted phases are used in more complex models, such as those imaging the core. Differential travel times between wave phases or types are also used.

10.8.3.1 Local Tomography

Local tomographic models are often based on a temporary seismic array targeting specific areas, unless in a seismically active region with extensive permanent network coverage. These allow for the imaging of the crust and upper mantle.

- Diffraction and wave equation tomography use the full waveform, rather than just the first arrival times. The inversion of amplitude and phases of all arrivals provide a more detailed density information than transmission travel time alone. Despite the theoretical appeal,

these methods are not widely employed because of the computing expense and difficult inversions.

- Reflection tomography originated with exploration geophysics. It uses an artificial source to resolve small-scale features at crustal depths. Wide-angle tomography is similar, but with a wide source to receiver offset. This allows for the detection of seismic waves refracted from subcrustal depths and can determine continental architecture and details of plate margins. These two methods are often used together.
- Local earthquake tomography is used in seismically active regions with sufficient seismometer coverage. Given the proximity between source and receivers, a precise earthquake focus location must be known. This requires the simultaneous iteration of both structure and focus locations in model calculations (Rawlinson et al., 2010).
- Teleseismic tomography uses waves from distant earthquakes that deflect upwards to a local seismic array. The models can reach depths similar to the array aperture, typically to depths for imaging the crust and lithosphere (a few hundred kilometres). The waves travel near 30° from vertical, creating a vertical distortion to compact features (Julian, 2006).

10.8.3.2 Regional or Global Tomography

Regional to global-scale tomographic models are generally based on long wavelengths. Various models have better agreement with each other than local models due to the large feature size they image, such as subducted slabs and superplumes. The trade-off from whole mantle to whole Earth coverage is the coarse resolution (hundreds of kilometres) and difficulty imaging small features (e.g. narrow plumes). Although often used to image different parts of the subsurface, *P* and *S* wave-derived models broadly agree where there is image overlap (Grand et al., 1997; ‘seismic tomography’, Wikipedia). These models use data from both permanent seismic stations and supplementary temporary arrays.

- First arrival travel time *P* wave data are used to generate the highest-resolution tomographic images of the mantle.
- In areas with limited seismograph or earthquake coverage, multiple phases of *S* waves can be used for tomographic models.
- Surface waves can be used for tomography of the crust and upper mantle, where no body wave (*P* and *S*) data are available.
- Attenuation tomography attempts to extract the anelastic signal from the elastic-dominated waveform of seismic waves.
- Ambient noise tomography cross-correlates waveforms from random wave fields generated by oceanic and atmospheric disturbances.
- Waveforms are modelled as rays in seismic analysis, but all waves are affected by the material near the ray path.

10.8.3.3 Applications of the Seismic Tomography

Seismic tomography can resolve anisotropy, inelasticity, density, and bulk sound velocity. Variations in these parameters may be a result of thermal or chemical differences, which are attributed to processes such as mantle plumes, subducting slabs, and mineral phase changes. Larger-scale features that can be imaged with tomography include the high velocities beneath continental shields and low velocities under ocean-spreading centres (Kearey et al., 2013).

- *Hotspots*. The mantle plume hypothesis proposes that areas of volcanism not readily explained by plate tectonics, called hotspots, are a result of thermal upwelling from as deep as the core–mantle boundary that become diapirs in the crust. This is an actively contested theory (Julian, 2006), although tomographic images suggest there are anomalies beneath

some hotspots. The best images of these are large low-shear-velocity provinces, or super-plume, visible on *S* wave models of the lower mantle and believed to reflect both thermal and compositional differences. The Yellowstone hotspot is responsible for volcanism at the Yellowstone Caldera and a series of extinct calderas along the Snake River Plain. The Yellowstone Geodynamic Project sought to image the plume beneath the hotspot (Smith et al., 2009). They found a strong low-velocity body from ~30 to 250 km depth beneath Yellowstone and a weaker anomaly from 250 to 650 km depth which dipped 60° west–north-west. The authors attribute these features to the mantle plume beneath the hotspot being deflected eastward by flow in the upper mantle seen in *S* wave models. The Hawaii hotspot produced the Hawaiian–Emperor seamount chain. Tomographic images show it to be 500 to 600 km wide and up to 2,000 km deep.

- *Subduction zones.* Subducting plates are colder than the mantle into which they are moving. This creates a fast anomaly that is visible in tomographic images. Both the Farallon plate that subducted beneath the West Coast of North America (Seismic Tomography, PDF, earthscope.org., Incorporated Research Institutions for Seismology (IRIS), retrieved 18 May 2016) and the northern portion of the Indian plate that has subducted beneath Asia have been imaged with tomography (Replumaz et al., 2010).

10.8.4 LIMITATIONS

Global seismic networks have expanded steadily since the 1960s but are still concentrated on continents and in seismically active regions. Oceans, particularly in the Southern Hemisphere, are under-covered (Julian, 2006). Tomographic models in these areas will improve when more data becomes available. The uneven distribution of earthquakes naturally biases models to better resolution in seismically active regions.

The type of wave used in a model limits the resolution it can achieve. Longer wavelengths are able to penetrate deeper into the Earth but can only be used to resolve large features. Finer resolution can be achieved with surface waves, with the trade-off that they cannot be used in models of the deep mantle. The disparity between wavelength and feature scale causes anomalies to appear of reduced magnitude and size in images. *P* and *S* wave models respond differently to the types of anomalies depending on the driving material property. First arrival time–based models naturally prefer faster pathways, causing models based on these data to have lower resolution of slow (often hot) features (Rawlinson et al., 2010). Shallow models must also consider the significant lateral velocity variations in continental crust.

Seismic tomography provides only the current velocity anomalies. Any prior structures are unknown, and the slow rates of movement in the subsurface (mm to cm per year) prohibit resolution of changes over modern timescales (Dziewonski, 2016). Tomographic solutions are non-unique. Although statistical methods can be used to analyze the validity of a model, unresolvable uncertainty remains. This contributes to difficulty comparing the validity of different model results. Computing power limits the amount of seismic data, number of unknowns, mesh size, and iterations in tomographic models. This is of particular importance in ocean basins, which, due to limited network coverage and earthquake density, require more complex processing of distant data. Shallow oceanic models also require smaller model mesh size due to the thinner crust (Liu and Gu, 2012).

Tomographic images are typically presented with a colour ramp representing the strength of the anomalies. This has the consequence of making equal changes appear of differing magnitude based on visual perceptions of colour, such as the change from orange to red being more subtle than blue to yellow. The degree of colour saturation can also visually skew interpretations. These factors should be considered when analyzing images (‘Seismic Tomography—Using earthquakes to image Earth’s interior’, Incorporated Research Institutions for Seismology (IRIS), retrieved 18 May 2016).

10.9 APPLICATIONS OF SEISMOLOGICAL DATA IN INDIA

10.9.1 SEISMOLOGICAL DATA OF THE SHILLONG PLATEAU AND THE ASSAM VALLEY BY THE GSI

A geophysical field investigation deploying seismological data analysis was carried out in the Shillong Plateau and Assam Valley area for deciphering the south-dipping hidden Oldham fault at the plateau and valley boundary (Saha et al., 2007). Several microearthquake surveys were carried out in the Shillong Plateau during the 1980s. It was reported that the Shillong Plateau earthquakes are mostly generated by a north-dipping reverse fault at the southern boundary of the plateau (Kayal, 1987; Kayal and De, 1991; Kayal, 2001). Kayal and Zhao (1998) studied seismic tomography using the temporary microearthquake network data and suggested that the north-dipping Dapsi thrust, the western extension of the Dauki fault, is the seismogenic fault. These earthquakes are mostly shallow (10–35 km), except for a few that were recorded from a depth range 40–50 km (Kayal, 2001). Chen and Molnar (1990), based on the teleseismic data, reported that maximum depth of an earthquake beneath the Shillong Plateau was 55 km. There were not much earthquake data available from the Assam Valley to the north of the plateau (Kayal and De, 1991). The alluvium-covered Assam Valley to the north was not well studied by the temporary networks in the 1980s.

Since 1990, the Assam Valley is, however, well covered by permanent telemetric and stand-alone digital microearthquake stations. These data are available in the form of yearly seismological bulletins jointly published by the National Geophysical Research Institute (NGRI) and the Regional Research Laboratory, Jorhat (RRL(J)). These bulletin data are reanalyzed by Bhattacharya et al. (2002), and the earthquakes are relocated by simultaneous inversion with much higher precision (Saha et al., 2007). A north–south depth section of the relocated events in the Shillong Plateau and Assam Valley along the 90.8°E longitude is shown in the figure in Saha et al. (2007). A careful examination of the section may suggest two reverse faults; one is north-dipping, which is comparable with the Dapsi thrust at the southern margin of the plateau, and the other (Oldham fault?) is south-dipping at the northern margin of the plateau (Saha et al., 2007). In addition to these, the section also suggests that the Brahmaputra fault may be a north-dipping seismogenic fault.

10.9.2 SEISMIC/EARTHQUAKE STUDIES, TRIPURA FOLD BELT, BY THE GSI

Seismic studies at GSI include macroseismic survey (postearthquake damage survey for assigning intensity), seismic hazard microzonation, microseismic (after-shock) survey, GPS monitoring for crustal movement, and seismic observatories. GSI focused in studying active fault (source region) and seismic microzonation of urban agglomeration in various geological domains of the country. Active fault mapping, including some interplate and intraplate faults, emphasizes the relevance of studying the nature of the source region of an earthquake.

In order to understand the seismicity pattern, nature of faulting, and subsurface heterogeneity in the entire Tripura fold belt and its adjoining areas, a comprehensive seismic monitoring was carried out by establishing a temporary seismic network consisting of nine seismographs (five broadband and four short-period) installed at different locations of the Tripura state. A total of 246 earthquakes (M_{5.0}) were located, and the dataset was plotted as the epicenter map shown (*Proc. Indian Natn. Sci. Acad.* 82 no. 3 July Spl. Issue 2016 pp. 1061–1081, DOI: 10.16943/ptinsa/2016/48504; tectonic map of Tripura fold belt showing the distribution of earthquakes; green triangles denote station locations, and solid circles locate earthquakes; data source: GSI, 2014). It was found that most of the earthquake magnitudes (M_L) are in the range of 2.0 to 4.0, and a few earthquakes are of magnitude more than 4.0. The focal depths of located earthquakes showed variation from 5 km to 150 km. Preliminary observation indicated a cluster of events in and around the Sylhet fault, orienting towards the NW direction, from the Khowai (KHW) seismograph station (GSI, 2014).

10.9.3 SEISMIC MICROZONATION OF PATNA TOWN BY THE GSI

The seismological and electrical signals are rigorously analyzed to estimate the seismogeophysical parameters to synthesize the plausible seismic microzonation map for the study area, with the aim of generating seismological parameters which have direct bearing on earthquake risk mitigation strategy to be adopted for the area under investigation. The peak frequency of the area varies between 0.58 Hz and 3.88 Hz (Central Headquarters Geological Survey of India, Kolkata, vol. 45, no.1, January–June 2014, Figure VII, p. 14, peak frequency [f_0] contour map [FS: 2012–2013 and 2013–14], GSI, 2014), while the peak frequency of the major part of the area lies below 1 Hz, with corresponding peak amplification factor varying between 2.19 and 11.97, as shown in the figure (Central Headquarters Geological Survey of India, Kolkata, vol. 45, no. 1, January–June 2014, Figure VIII, p. 15, Peak Amplification Factor [A0] Contour Map [FS: 2012–2013 and 2013–14], GSI, 2014). It is observed that the higher values of amplification factors are prevalent at very few locations that may be due to in-filled area or due to shallow water level or due to surface sand layer or a combination of these factors. The presence of the filled palaeochannel beneath the study area may have sufficient potential to amplify the seismic signals of the subsurface material properties associated with the palaeochannel. The high-conductivity and low-resistivity values of the subsurface layers were found well corroborative with water-saturated in-filled subsurface formations for the area under investigation.

REFERENCES

- Bhattacharya, P.M., Majumdar, R.K. and Kayal, J.R., 2002. Fractal dimension and b-value mapping in north-east India. *Current Science*, vol. 82, pp. 1486–1491.
- Chen, W.P. and Molnar, P., 1990. Source parameters of earthquakes and intraplate deformation beneath the Shillong Plateau and northern Indo-Burma region. *Journal of Geophysical Research*, vol. 95, pp. 527–552.
- Chinenyeze, M., 2015. Types of seismic energy sources for petroleum exploration in desert, dry-land, swamp and marine environments in Nigeria and other sub-saharan Africa. *International Journal of Science and Research (IJSR)*, vol. 6.
- Dziewonski, A., 2016. Global seismic tomography: What we really can say and what we make up (PDF). Available at mantleplumes.org.
- Gisiner, R.C., 2016. Sound and marine seismic surveys. *Acoustical Society of America: Acoustic Today*, vol. 12, no. 4, pp. 10–18.
- Grand, S.P., Hilst, V.R.D. and Widiyantoro, S., 1997. Global seismic tomography: A snapshot of convection in the earth. *GSA Today*, vol. 7, pp. 1–7.
- GSI, 2014. Scientific news: A glimpse. *Central Head Quarters (CHQ) e-News*, vol. 45, no. 1, pp. 1–68.
- James, R.B., Lauderdale, W.W. and Harold, K., 1979. *ALSEP Termination Report*. Washington, DC: National Aeronautics and Space Administration (NASA), Scientific and Technical Information Office. Reference Publication 1036.
- Julian, B., 2006. Seismology: The hunt for plumes (PDF). Available at mantleplumes.org.
- Kayal, J.R., 1987. Microseismicity and source mechanism study: Shillong Plateau, northeast India. *Bulletin of the Seismological Society of America (BSSA)*, vol. 77, pp. 184–194.
- Kayal, J.R., 2001. Microearthquake activity in some parts of Himalaya and the tectonic model. *Tectonophysics*, vol. 339, pp. 331–351.
- Kayal, J.R. and De, R., 1991. Microseismicity and tectonics in northeast India. *Bulletin of the Seismological Society of America (BSSA)*, vol. 81, pp. 131–138.
- Kayal, J.R. and Zhao, D., 1998. Three dimensional seismic structure beneath Shillong Plateau and Assam Valley, northeast India. *Bulletin of the Seismological Society of America (BSSA)*, vol. 88, pp. 667–676.
- Kearey, P. and Brooks, M., 2002. *An Introduction to Geophysical Exploration*. Oxford: Blackwell Science, p. 288.
- Kearey, P., Klepeis, K.A. and Frederick J.V., 2013. *Global Tectonics*. Hoboken, NJ: John Wiley & Sons, ISBN 9781118688083.
- Kirsch, R., 2006. Umweltgeophysik in der Praxis: Untersuchung von Alttablagerungen und kontaminierten Standorten, Script. In: Kirsch, R. (ed), *Groundwater Geophysics*. Berlin, Heidelberg: Springer, DOI: 10.1007/3-540-29387-6_17, Print ISBN 9783540293835, Online ISBN 9783540293873.

- Knapp, R.W. and Steeples, D.W., 1986b. High-resolution common-depth-point reflection profiling—field acquisition parameter design. *Geophysics*, vol. 51, pp. 283–294.
- Knödel, K., Krummel, H. and Lange, G., 1997. *Geophysik.—Handbuch zur Erkundung des Untergrundes von Deponien und Altlasten Band 3*. Berlin, Heidelberg: Springer.
- Lavergne, M., 1989. *Seismic Methods*. Paris: Éditions Technip, ISBN: 2-7108-0552-9.
- Liu, Q. and Gu, Y.J., 2012. Seismic imaging: From classical to adjoint tomography. *Tectonophysics*, pp. 566–567, 31–66. DOI: 10.1016/j.tecto.2012.07.006.
- Lowrie, W., 2007. *Fundamentals of Geophysics*. 2nd edn. New York: Cambridge University Press, p. 375.
- Madu, A.J.C., Eze, C.L. and Otuokere, I.E., 2017. Investigation of possible impact of 2, 4, 6-Trinitrotoluene (TNT) explosive seismic energy source on the nitrate content of groundwater in Sagbama Area, Niger Delta, Nigeria. *Research Journal of Engineering and Technology*, vol. 8, no. 1, p. 63, DOI: 10.5958/2321-581x.2017.00010.1, ISSN 0976-2973.
- Mari, J.L. and Porel, G., 2018. Contribution of seismic and acoustic methods to the characterization of karstic formation. In: Mari, J.L., Vergniault, C. (eds), *Well Seismic Surveying and Acoustic Logging, EDP Sciences*, ISBN (e-book) 7598-2263-8, DOI: 10.10051/978-2-7598-2263-8, Available at www.edp-open.org/wellseismic-surveying-and-acoustic-logging.
- Nolet, G., 1987. Seismic wave propagation and seismic tomography. In: Nolet, G. (ed), *Seismic Tomography. Seismology and Exploration Geophysics*. Netherlands: Springer, pp. 1–23, DOI: 10.1007/978-94-009-3899-1_1, ISBN 9789027725837.
- Parkes, G.E. and Hatton, L., 1986. *The Marine Seismic Source*. 1st edn. Dordrecht, the Netherlands: Springer Science Business Media, DOI: 10.1007/978-94-017-3385-4.
- Pelton, J.R., 2005. Near-surface seismology: Surface-based methods. In: Butler, D.K. (ed), *Near-Surface Geophysics*. Tulsa, OK: Society of Exploration Geophysicists, pp. 219–264.
- Rabbel, W., 2006. Seismic methods. In: Kirsch, R. (ed), *Groundwater Geophysics—A Tool for Hydrogeology*. Berlin, Heidelberg: Springer, pp. 23–84.
- Rawlinson, N., Pozgay, S. and Fishwick, S., 2010. Seismic tomography: A window into deep Earth. *Physics of the Earth and Planetary Interiors*, vol. 178, no. 3–4, pp. 101–135. DOI: 10.1016/j.pepi.2009.10.002.
- Replumaz, A., Negredo, A.M., Guillot, S. and Villaseñor, A., 2010. Multiple episodes of continental subduction during India/Asia convergence: Insight from seismic tomography and tectonic reconstruction. *Tectonophysics: Convergent Plate Margin Dynamics: New Perspectives from Structural Geology, Geophysics and Geodynamic Modelling*, vol. 483, no. 1–2, pp. 125–134. DOI: 10.1016/j.tecto.2009.10.007.
- Reynolds, J.M., 1998. *An Introduction to Applied and Environmental Geophysics*. Chichester and New York: John Wiley.
- Romanowicz, B., 2003. Global mantle tomography: Progress status in the past 10 years. *Annual Review of Earth and Planetary Sciences*, vol. 31, no. 1, pp. 303–328. DOI: 10.1146/annurev.earth.31.091602.113555.
- Saha, D.K., Naskar, D.C., Bhattacharya, P.M. and Kayal, J.R., 2007. Geophysical investigation for the hidden Oldham fault: Magnetotelluric and Deep resistivity surveys in Shillong Plateau and Assam valley. *Geological Society of India*, vol. 69, pp. 359–372.
- Shearer, P., 2016. *Introduction to Seismology*. United States Geologic Survey (USGS). New York: Cambridge University Press.
- Sheriff, R.E., 2002a. *Encyclopedic Dictionary of Applied Geophysics (Geophysical References No. 13)*. 4th edn. Houston, TX: GeoScienceWorld, Society of Exploration Geophysicists Geophysical References Series, p. 429, ISBN 9781560801184.
- Sheriff, R.E., 2002b. *Encyclopaedic Dictionary of Exploration Geophysics*. 4th edn. Tulsa, OK: Society of Exploration Geophysicists, p. 357.
- Sheriff, R.E. and Geldart, L.P., 1995. *Exploration Seismology*. 2nd edn. New York: Cambridge University Press.
- Smith, R.B., Jordan, M., Steinberger, B., Puskas, C.M., Farrell, J., Waite, G.P., Husen, S., Chang, W.L. and O'Connell, R., 2009. Geodynamics of the Yellowstone hotspot and mantle plume: Seismic and GPS imaging, kinematics, and mantle flow. *Journal of Volcanology and Geothermal Research. The Track of the Yellowstone Hotspot What do Neotectonics, Climate Indicators, Volcanism, and Petrogenesis Reveal about Subsurface Processes?*, vol. 188, no. 1–3, pp. 26–56. DOI: 10.1016/j.jvolgeores.2009.08.020.
- Stark, A., 2010. *Seismic Methods and Applications*. Illustrated edn. Irvine, CA: Universal-Publishers and Brown Walker Press, pp. 177–178; p. 592, ISBN 9781599424439.

- Steeple, D.W., 2005. Shallow seismic methods. In: Rubin, Y, Hubbard, S.S. (eds), *Hydrogeophysics*. Dordrecht, The Netherlands: Springer, pp. 215–251.
- Strobbia, C., Vermeer, P., Glushchenko, A. and Laake, A., 2009. Advances in surface-wave processing for nearsurface characterisation in land seismic. In: 71st EAGE Conference and Exhibition—Workshops and Fieldtrips. Netherlands: EAGE Publications BV, DOI: 10.3997/2214-4609.201404894, ISBN 9789462821033.
- Telford, W.M., Geldart, L.P. and Sheriff, R.E., 1990. *Applied Geophysics*. Cambridge: Cambridge University Press.
- Telford, W.M., Geldart, L.P. and Sheriff, R.E., 1991. *Applied Geophysics*. 2nd edn. Cambridge: Cambridge University Press, p. 770, Hard cover ISBN 052132693.
- Yilmaz, O., 2001. *Seismic Data Analysis*, vol. 1 and 2, 2nd edn. Tulsa, OK: Society of Exploration Geophysicists.
- Yordkayhun, S., Ivanova, A., Giese, R., Juhlin, C. and Cosma, C., 2009. Comparison of surface seismic sources at the CO2SINK site, Ketzin, Germany. *Geophysical Prospecting*, vol. 57, no. 1, pp. 125–139. DOI: 10.1111/j.1365-2478.2008.00737.x, ISSN 0016-8025.

11 Radioactivity Method

11.1 INTRODUCTION

The geophysical techniques described in previous chapters have strong dependence and one-to-one correspondence to changes of physical property contrasts associated with the gravity, magnetic, electrical, or electromagnetic properties of geological formations of subsurface strata consisting of various types of rocks and minerals. In the mid-20s, another property of certain elements has become a challenge to decipher for the sake of its considerable economic importance. This property is known as the radioactive property of rocks and minerals. *Radioactivity* is a phenomenon by which the radioactive elements associated with rocks and minerals are subjected to emanate energy through the disintegration of the atomic nuclei of radioactive elements from host rocks by spontaneous emissions of radiation. In this chapter, we have sufficiently explained the fundamentals of radioactive processes so that diverse applications of the radioactive method can be understood in the sense of how the method works as a tool for the geophysical exploration of minerals, and its further applications for better understanding the crustal processes, by analyzing radon and helium emissions from the source zones of earthquakes in seismically active regions of the world.

It is known to us that the original discovery was made by Becquerel in 1896 (Becquerel, 1896), which happened just shortly after Roentgen had announced in 1895 the discovery of X-rays. Becquerel found that minerals containing uranium, as well as salts of uranium, emitted radiations that passed through material opaque to ordinary light, affected photographic emulsions in a manner similar to X-rays, and would ionize a gas.

The discovery of other radioactive elements soon followed. Mme Curie investigated minerals of uranium and extracted two new elements, polonium and radium, which were much more active than uranium. About the same time, Schmidt discovered that thorium was radioactive, and Debierne found the new radioactive element actinium.

With the advent of advanced analytical and instrumentation capabilities, more than 20 naturally occurring elements are now known to be radioactive. With reference to geophysical exploration, it is observed that uranium (U), thorium (Th), and an isotope of potassium (K) are of much importance in exploration. Two others, rubidium and isotopes of carbon, are being used for determining the ages of rocks, but the rest are either so rare or have feeble radioactive characteristics, so weakly radioactive, or both, which will have no significance in the field of applied geophysics as to be of no significance in applied geophysics. These two elements, uranium and thorium, are important today as a source of nuclear fuel for the generation of heat and power in nuclear reactors.

11.1.1 RADIOACTIVITY: A NATURAL PROPERTY OF CERTAIN ATOMS

In nature, the nuclei of most atoms are stable. However, certain atoms have unstable nuclei due to an excess of either protons or neutrons or an excess of both. They are described as radioactive and are known as radioisotopes or radionuclides (CEA, 2005, Communication Division: 'From radioelements to scientific applications'; www.cea.fr. ISSN 1637-5408). The nuclei of radioactive atoms change spontaneously into other atomic nuclei, which may or may not be radioactive. For instance, uranium-238 changes into a succession of different radioactive nuclei until it reaches a stable form, lead (Pb)-206. This irreversible transformation of a radioactive atom into a different type of atom is known as disintegration. It is accompanied by the emission of different types of radiation. A chemical element can therefore have both radioactive isotopes

and nonradioactive isotopes, which can be exemplified by carbon-12 as not radioactive, but carbon-14 is a radioactive isotope of carbon. Because radioactivity only affects the nucleus and not the electrons, the chemical properties of radioactive isotopes are the same as those of stable isotopes. Once atoms have the same number of protons and different number of neutrons, then they belong to the same chemical element, as for example, carbon-12 (six neutrons) and carbon-14 (eight neutrons) are two carbon isotopes. The chemical properties of an atom are determined by the number of electrons.

11.2 RADIOACTIVE DECAY AND HALF-LIFE

An unstable nucleus will decompose spontaneously, or decay, into a more stable configuration but will do so only in a few specific ways, by emitting certain particles or certain forms of electromagnetic energy. *Radioactive decay* is a property of several naturally occurring elements as well as of artificially produced isotopes of the elements. The radioactive disintegration of a particular nucleus is a random phenomenon (OpenStax CNX: cnx.org). *Radioactivity* is measured as the number of disintegrations per second within a sample. The product of a radioactive decay process, called the daughter of the parent isotope, may itself be unstable, in which case it, too, will decay. The process continues until a stable nuclide has been formed. The activity of a radioactive sample diminishes over time, with the gradual disappearance of the unstable nuclei it contains. However, for each radioactive isotope, it is possible to give a half-life, which is the time needed for half of the radioactive atoms present at the outset to disappear by spontaneous transformation. Depending on the nucleus, radioactivity can last a few seconds, several days, or billions of years. Half-lives range from more than 10^9 years for some nuclei to less than 10^{-9} second (Table 11.1). Consequently, the number of nuclei, dN , which decay in a short time, dt , is directly proportional to the number of nuclei, N , present. That is:

$$\frac{dN}{dt} = -\lambda N \quad (11.1)$$

where λ is decay constant. Integrating eq. 11.1 leads to

$$N = N_0 e^{-\lambda t} \quad (11.2)$$

The half-life $t_{\frac{1}{2}}$ is the time during which half of the radioactive nuclei decay gets decayed, that is, when $N = \frac{1}{2} N_0$. Therefore, from eq. 11.2 it follows that

$$t_{\frac{1}{2}} = \frac{0.693}{\lambda}$$

In order to calculate the half-life of a radioactive material, we use the following equation:

$$N(t) = N_0 e^{-\left(\frac{0.693 t}{t_{\frac{1}{2}}}\right)}$$

where N_0 is the initial quantity of the substance, $N(t)$ the quantity that still remains and has not yet decayed after a time t , $t_{\frac{1}{2}}$ the half-life of the decaying quantity, and e the Euler's number, which equals 2.71828.

For example, cobalt (Co)-60, an isotope that emits gamma rays used to treat cancer, has a half-life of 5.27 years (for cobalt-60, which has a half-life of 5.27 years, 50% remains after 5.27 years

TABLE 11.1
Half-Lives of a Number of Radioactive Bodies

Chemical Elements	Radioactive Half-Life	Origin	Where Present	Examples of Use
Tritium	12.3 yrs	Artificial	—	Thermonuclear fusion, biological tagging
Carbon-11	20.4 min	Artificial	—	Medical imaging
Carbon-14	5,730 yrs	Natural	Atmosphere carbon compounds	Dating
Oxygen-15	2.02 mins	Artificial	—	Medical imaging
Phosphorus-32	14.3 days	Artificial	—	Biological research
Sulphur-35	87.4 days	Artificial	—	Biological research
Potassium-40	1.3 billion yrs	Natural	Rocks rich in potassium, skeleton	—
Cobalt-60	5.27 yrs	Artificial	—	Radiotherapy, industrial irradiation, gamma radiography
Strontium-90	28.8 yrs	Artificial	Produced by nuclear reactors	Thickness gauges
Iodine-123	13.2 hrs	Artificial	—	Nuclear medicine
Iodine-131	8.05 days	Artificial	Produced by nuclear reactors	Nuclear medicine therapy
Cesium-137	30.2 yrs	Artificial	Produced by nuclear reactors	Brachytherapy
Thallium-201	3.04 days	Artificial	—	Nuclear medicine
Radon-222	3.82 days	Natural	Gas released by granite rocks	—
Radium-226	1,600 yrs	Natural	Rock containing uranium	—
Thorium-232	14 billion yrs/giga-annum (10^9 yrs)	Natural	—	Mineral dating
Uranium-235	704 million yrs (10^6 yrs)	Natural	Some terrestrial rock/granite rock	Nuclear deterrent fuel
Plutonium-239	24,100 yrs	Artificial	Produced by nuclear reactors	Nuclear deterrent fuel

Source: Adapted from CEA (2005), Communication Division: 'From radioelements to scientific applications' (www.cea.fr): ISSN 1637–5408.

[one half-life], 25% remains after 10.54 years [two half-lives], 12.5% remains after 15.81 years [three half-lives], and so on) (OpenStax; www.cnx.org). In a given cobalt-60 source, since half of the $^{60}_{27}\text{Co}$ nuclei decay every 5.27 years, it may be noted that for a given substance, the intensity of radiation that it produces is directly proportional to the rate of decay of the substance and the amount of the substance. This is as expected for a process following first-order kinetics. Thus, a cobalt-60 source that is being used for cancer treatment must be replaced regularly to continue to be effective.

11.2.1 THE NATURE OF RADIOACTIVE EMISSIONS

The emissions of the most common forms of spontaneous radioactive decay are the alpha (α) particle, the beta (β) particle, the gamma (γ) ray, and the neutrino. The alpha particle is actually the nucleus of a helium-4 atom, with two positive charges ^4_2He . Such charged atoms are called ions. The neutral helium atom has two electrons outside its nucleus balancing these two charges. Beta particles may be negatively charged (beta-minus, symbol β^-) or positively charged (beta-plus, symbol β^+). The beta-minus (β^-) particle is actually an electron created in the nucleus during beta decay

without any relationship to the orbital electron cloud of the atom. The beta-plus particle, also called the positron, is the antiparticle of the electron. When brought together, two such particles will mutually annihilate each other. Gamma rays are electromagnetic radiations, such as radio waves, light, and X-rays. Beta radioactivity also produces neutrino and antineutrino, particles that have no charge and very little mass, symbolized by ν and $\bar{\nu}$, respectively.

In the less-common forms of radioactivity, fission fragments, neutrons, or protons may be emitted. Fission fragments are themselves complex nuclei with usually between one-third and two-thirds the charge Z and the atomic mass A of the parent nucleus. Neutrons and protons are, of course, the basic building blocks of complex nuclei, having approximately unit mass on the atomic scale and having zero charge or unit positive charge, respectively. The neutron cannot survive for long in the free state. It is rapidly captured by nuclei in matter; otherwise, in free space, it will undergo beta-minus decay to a proton, an electron, and an antineutrino with a half-life of 12.8 minutes. The proton is the nucleus of ordinary hydrogen and is stable.

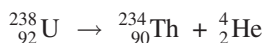
11.2.2 TYPES OF RADIOACTIVITY

Ernest Rutherford's experiments involving the interaction of radiation with a magnetic or electric field (Figure 11.1) helped him determine that one type of radiation consisted of positively charged and relatively massive α particles, a second type was made up of negatively charged and much less massive β particles, and a third were uncharged electromagnetic waves, γ rays. We now know that α particles are high-energy helium nuclei, β particles are high-energy electrons, and γ radiation is composed of high-energy electromagnetic radiation. We classify different types of radioactive decay by the radiation produced.

11.2.2.1 Alpha Decay

In alpha decay, an energetic helium ion (alpha particle) is ejected, leaving a daughter nucleus of atomic number two less than the parent's and of atomic mass number four less than the parent's. An example is the decay (symbolized by an arrow) of the abundant isotope of uranium, ^{238}U , to a thorium daughter plus an alpha particle:

$$Q_{\alpha} = 4.268 \text{ MeV}$$



$$t_{\frac{1}{2}} = 4.51 \times 10^9 \text{ years}$$

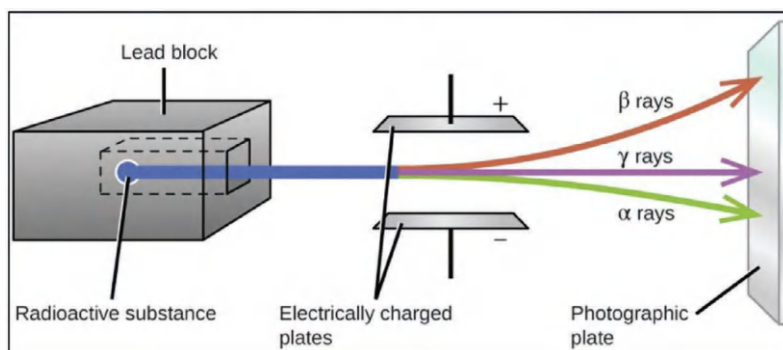


FIGURE 11.1 Alpha particles, which are attracted to the negative plate and deflected by a relatively small amount, must be positively charged and relatively massive. Beta particles, which are attracted to the positive plate and deflected a relatively large amount, must be negatively charged and relatively light. Gamma rays, which are unaffected by the electric field, must be uncharged.

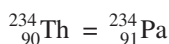
Source: <https://opentextbc.ca/chemistry/chapter/21-3-radioactive-decay/>.

Given for this and subsequent reactions are the energy released (Q) in millions of electron volts (MeV) and the half-life ($t_{\frac{1}{2}}$). It should be noted that in alpha decays, the charges, or number of protons, shown in subscript are in balance on both sides of the arrow, as are the atomic masses, shown in superscript.

11.2.2.2 Beta-Minus Decay

In beta-minus decay, an energetic negative electron is emitted, producing a daughter nucleus of one higher atomic number and the same mass number. An example is the decay of the uranium daughter product thorium-234 into protactinium-234:

$$Q_{\beta^-} = 0.263 \text{ MeV}$$



$$t_{\frac{1}{2}} = 24.1 \text{ days}$$

Here, the number of protons is increased by one in the reaction, but the total charge remains the same, because an electron, with a negative charge, is also created. This can be explained upon taking the following example with the supposition that we are having a bunch of cobalt-60 atoms. Cobalt-60 decays down to nickel-60 by beta emission. In this context, we have to know the time of decay of the cobalt-60 in sense of when it gets decayed. The second point is to know the stages of decay either once at a time or randomly. Initially, it may seem like atoms decay randomly, but their probability of decaying can be predicted by knowing an atom's half-life. It turns out that the half-life of cobalt-60 is about 5.27 years.



11.2.2.3 Gamma Decay

A third type of radiation, gamma radiation, is usually accompanied with alpha or beta decay. Gamma rays are photons and are without rest mass or charge. Alpha or beta decay may simply proceed directly to the ground (lowest energy) state of the daughter nucleus without gamma emission, but the decay may also proceed wholly or partly to higher energy states (excited states) of the daughter. In the latter case, gamma emission may occur as the excited states transform to lower-energy states of the same nucleus (Figure 11.2).

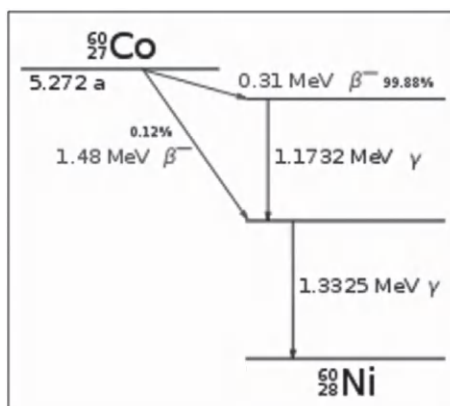


FIGURE 11.2 Gamma emission spectrum of cobalt.

Source: Wikipedia.

An alternative explanation has also been provided for gamma emission, which occurs when an excited nucleus gets transformed to a lower energy state by ejecting an electron from the cloud surrounding the nucleus. This orbital electron ejection is known as internal conversion and gives rise to an energetic electron and often an X-ray as the atomic cloud fills in the empty orbital of the ejected electron. The ratio of internal conversion to the alternative gamma emission is called the internal conversion coefficient.

11.2.2.3.1 Isomeric Transition

There is a wide range of rates of half-lives for the gamma emission processes. Usually, in dipole transitions, as explained under gamma transition, the gamma ray carries off one h unit of angular momentum, which is the fastest having time of emission, lesser than nanoseconds ($1 \text{ ns} = 10^{-9} \text{ sec}$). The law of conservation of angular momentum requires that the sum of the angular momenta of the radiation and the daughter nucleus is equal to the angular momentum (spin) of the parent. If the spins of initial and final states differ by more than one, then dipole radiation is forbidden, whilst gamma emission shall occur more slowly by a higher-multipole (quadrupole, octupole, etc.) gamma transition. If the gamma emission half-life exceeds about 1 ns, the excited nucleus is said to be in a metastable, or isomeric, state that indicates the names for a long-lived excited state, and it is customary to classify the decay as another type of radioactivity, an isomeric transition. An example of isomerism is found in the protactinium-234 nucleus of the uranium-238 decay chain:

$$Q_\gamma = 0.0698 \text{ MeV}$$

$${}^{234m}_{91}\text{Pa} = {}^{234}_{91}\text{Pa} + \gamma$$

$$t_{\frac{1}{2}} = 1.17 \text{ min}$$

The letter m following the mass number stands for ‘metastable’ and indicates a nuclear isomer of protactinium (Pa).

11.2.2.4 Beta-Plus Decay vis-à-vis Positron Emission

During the 1930s, new types of radioactivity were found among the artificial products of nuclear reactions: beta-plus decay, or positron emission, and electron capture. In beta-plus decay, an energetic positron is created and emitted, along with a neutrino, and the nucleus transforms to a daughter, lower by one in atomic number and the same in mass number. For instance, carbon-11 ($Z = 6$) decays to boron-11 ($Z = 5$), plus one positron and one neutrino:

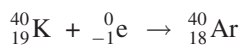
$$Q_{\beta+} = 0.97 \text{ MeV}$$

$${}^{11}_6\text{C} = {}^{11}_5\text{B} + e^+ + \nu$$

$$t_{\frac{1}{2}} = 20.4 \text{ min}$$

11.2.2.5 Electron Capture

Electron capture occurs when one of the inner electrons in an atom is captured by the atom’s nucleus. For example, potassium (K)–40 undergoes electron capture:



Electron capture occurs when an inner shell electron combines with a proton and is converted into a neutron. The loss of an inner shell electron leaves a vacancy that will be filled by one of the

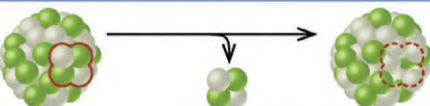
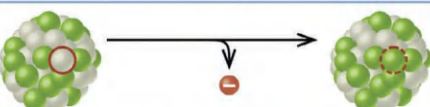
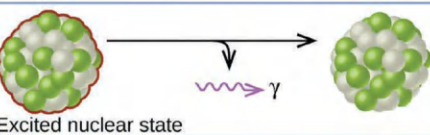
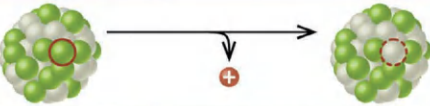
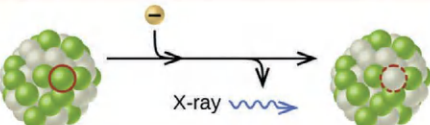
Type	Nuclear equation	Representation	Change in mass/atomic numbers
Alpha decay	${}^A_ZX \rightarrow {}^4_2\text{He} + {}^{A-4}_{Z-2}Y$		A: decrease by 4 Z: decrease by 2
Beta decay	${}^A_ZX \rightarrow {}^0_{-1}e + {}^{A}_{Z+1}Y$		A: unchanged Z: increase by 1
Gamma decay	${}^A_ZX \rightarrow {}^0_0\gamma + {}^A_ZY$		A: unchanged Z: unchanged
Positron emission	${}^A_ZX \rightarrow {}^0_{+1}e + {}^{A}_{Z-1}Y$		A: unchanged Z: decrease by 1
Electron capture	${}^A_ZX + {}^0_{-1}e \rightarrow {}^{A}_{Z-1}Y + \gamma$		A: unchanged Z: decrease by 1

FIGURE 11.3 This table summarizes the type, nuclear equation, representation, and any changes in the mass or atomic numbers for various types of decay.

Source: <https://opentextbc.ca>.

outer electrons. As the outer electron drops into the vacancy, it will emit energy. In most cases, the energy emitted will be in the form of an X-ray. Like positron emission, electron capture occurs for ‘proton-rich’ nuclei that lie below the band of stability. Electron capture has the same effect on the nucleus as happens with positron emission: the atomic number is decreased by one, and the mass number does not change. This increases the neutron–proton ratio, and the daughter nuclide lies closer to the band of stability than did the parent nuclide. Whether electron capture or positron emission occurs is difficult to predict. The choice is primarily due to kinetic factors, with the one requiring the smaller activation energy being the one more likely to occur. Figure 11.3 summarizes these types of decay, along with their equations and changes in atomic and mass numbers.

11.2.3 RADIOACTIVE DECAY SERIES

The naturally occurring radioactive isotopes of the heaviest elements fall into chains of successive disintegrations, or decays, and all the species in one chain constitute a radioactive family, or radioactive decay series. Three of these series include most of the naturally radioactive elements of the periodic table. They are the uranium (U) series, the actinide (Ac) series, and the thorium (Th) series. The neptunium (Np) series is a fourth series, which is no longer significant on the Earth because of the short half-lives of the species involved. Each series is characterized by a parent (first member) that has a long half-life and a series of daughter nuclides that ultimately lead to a stable end product: a nuclide on the band of stability (Figure 11.4). In all three series, the end product is a stable isotope of lead. The neptunium series, previously thought to terminate with bismuth-209, terminates with thallium (Tl)-205.

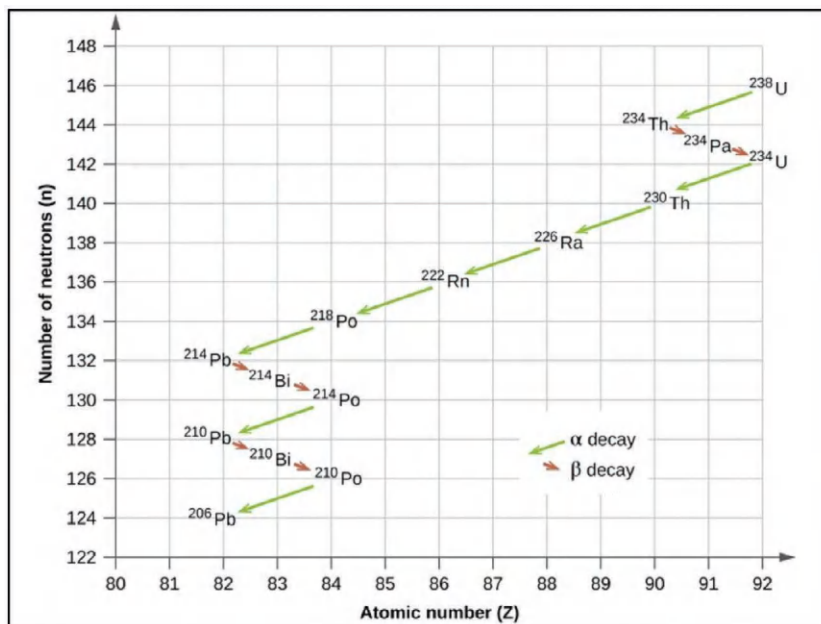


FIGURE 11.4 Uranium-238 undergoes a radioactive decay series consisting of 14 separate steps before producing stable lead-206. This series consists of eight α decays and six β decays.

Source: <https://opentextbc.ca/chemistry/chapter/21-3-radioactive-decay/>.

11.2.4 PENETRATION BY α , β , AND γ RAYS

Alpha, beta, and gamma rays produce three different effects in varying degrees, which forms the basis of the radioactive method. They affect photographic emulsions in the same way as light and X-rays. They ionize gas, make it electrically conducting. They scintillate or phosphorescence in certain minerals and chemical compounds (Figure: Penetration by α , β and γ rays, adapted from Weisstein, 2014).

1. *Alpha radiation* is a stream of alpha particles (equivalent to helium nucleus, charge-positive, tightly bound entity within nuclei heavier than helium), which are positively charged. These are fairly large, which means these have a difficult time getting through materials like clothes and paper. Alpha particles produce very intense ionization in gases and travel with velocities ranging from 1.4 to 2.2×10^{10} cm/sec.
2. *Beta radiation* is a stream of beta particles, which are negatively charged. These particles can more easily penetrate materials like clothes, and they can even get deep into your skin, where they can do harm to your cells. But these can be blocked with denser materials like aluminium. Beta particles, which are electrons, are emitted with velocities ranging from 10^{10} cm/sec to nearly the velocity of light (3×10^{10} cm/sec).
3. *Gamma radiation* is high-frequency electromagnetic radiation. Gamma rays have no charge but have a lot of energy—more energy than even visible light or X-rays. Because of this, they are able to pass right through most materials, making them quite dangerous. But they cannot penetrate into very dense materials like lead, which is why we generally find a lead vest being used in a hospital or laboratory for protection from the worst consequences due to emissions of radioactive rays.

11.3 UNITS OF RADIOACTIVITY

The International System of Units (SI) unit of radioactive activity is the becquerel (Bq), named in honour of the scientist Henri Becquerel: 1 Bq is defined as one transformation (or decay or disintegration) per second. An older unit of radioactivity is the curie, C_i , which was originally defined as ‘the quantity or mass of radium emanation in equilibrium with one gram of radium (element)’. Today, the curie is defined as 3.7×10^{10} disintegrations per second, so that $1 C_i = 3.7 \times 10^{10}$ Bq. The effects of ionizing radiation are often measured in units of gray for mechanical or sievert for damage to tissue.

11.4 RADIOACTIVE DETECTOR

The exploration for radioactive minerals utilizing various instruments, generally a Geiger counter or scintillation counter, by measuring the natural radioactivity of Earth materials.

11.4.1 GEIGER–MULLER (GM) COUNTER

A *Geiger counter* is an instrument used for detecting and measuring ionizing radiation. Also known as a Geiger–Müller counter (or Geiger–Müller counter), it is widely used in applications such as radiation dosimetry, radiological protection, experimental physics, and the nuclear industry. The original detection principle was realized in 1908 at the Victoria University of Manchester, but it was not until the development of the Geiger–Müller tube in 1928 (Rutherford and Geiger, 1908) that the Geiger counter could be produced as a practical instrument. Since then, it has been very popular due to its robust sensing element and relatively low cost. However, there are limitations in measuring high radiation rates and the energy of incident radiation (Glenn, 2000). The GM counter detects ionizing radiation such as alpha particles, beta particles, and gamma rays using the ionization effect produced in a Geiger–Müller tube, which gives its name to the instrument. In wide and prominent use as a handheld radiation survey instrument, it is perhaps one of the world’s best-known radiation-detection instruments.

11.4.1.1 Operating Principle of the GM Counter

A Geiger counter consists of a Geiger–Müller tube that senses an element, which detects the radiation and the processing electronics, then displays the result. The Geiger–Müller tube is filled with an inert gas, such as helium, neon, or argon, at low pressure, to which a high voltage is applied. The tube briefly conducts an electrical charge when a particle or photon of incident radiation makes the gas conductive by ionization. The ionization is considerably amplified within the tube by the Townsend effect (Townsend, 1901) to produce an easily measured detection pulse which is fed to the processing and display electronics. This large pulse from the tube makes the Geiger counter relatively cheap to manufacture, as the subsequent electronics are greatly simplified (Glenn, 2000). The electronics also generate the high voltage, typically 400–900 V, that has to be applied to the Geiger–Müller tube to enable its operation. To stop the discharge in the Geiger–Müller tube, a little halogen gas or organic material (alcohol) is added to the gas mixture.

There are two types of detected radiation readout consisting of counts or radiation dose. The counts display is the simplest and is the number of ionizing events detected, displayed either as a count rate, such as ‘counts per minute’ or ‘counts per second’, or as a total number of counts indicating an integrated total over a set time period. The counts readout is normally used when alpha or beta particles are being detected. More complex to achieve is a display of radiation dose rate, displayed in a unit such as the ‘sievert’, which is normally used for measuring gamma or X-ray dose rates. A Geiger–Müller tube can detect the presence of radiation, but not its energy, which influences the radiation’s ionizing effect. Consequently, instruments measuring dose rate require the

use of an energy-compensated Geiger–Müller tube so that the dose displayed relates to the counts detected. The electronics will apply known factors to make this conversion, which is specific to each instrument and is determined by design and calibration. The readout can be analogue or digital, and modern instruments offer serial communications with a host computer or network.

There are two main limitations of the Geiger counter. Because the output pulse from a Geiger–Müller tube is always of the same magnitude (regardless of the energy of the incident radiation), the tube cannot differentiate between radiation types. Secondly, the inability to measure high radiation rates due to the ‘dead time’ of the tube. This is an insensitive period after the ionization of the gas occurs, during which any further incident radiation will not result in a count, and the indicated rate is, therefore, lower than actual. Typically, the dead time will reduce indicated count rates above about 10^4 to 10^5 counts per second, depending on the characteristic of the tube being used for the purpose. In contrary to that, some counters have circuitry which can compensate for this, making an accurate measurement of ion, for which chamber instruments are preferred for high radiation rates.

11.4.1.2 Types and Applications

The intended detection application of a Geiger counter dictates the tube design used. Consequently, there are a great many designs, but they can be generally categorized as ‘end-window’, ‘window-less’, ‘thin-walled’, ‘thick-walled’, and sometimes hybrids of these types.

The first historical uses of the Geiger principle were for the detection of alpha and beta particles, and the instrument is still used for this purpose today. For alpha particles and low-energy beta particles, the ‘end-window’ type of a Geiger–Müller tube has to be used as these particles have a limited range and are easily stopped by a solid material. Therefore, the tube requires a window which is thin enough to allow as many as possible of these particles through to the fill gas. The window is usually made of mica.

Gamma and X-ray detections, collectively known as photons, are made by Geiger counters associated with the windowless tube under its wider applicability. However, detection efficiency is low compared to that of alpha and beta particles. The article on the Geiger–Müller tube carries a more detailed account of the techniques used to detect photon radiation. It is well documented that in case of high-energy photons, the tube relies on the interaction of the radiation with the tube wall, usually a high-Z material, such as chrome steel of 1–2 mm thickness, to produce electrons within the tube wall. These enter and ionize the fill gas.

Neutron detection is also made through the use of the Geiger tube, where the gas used is boron trifluoride or helium-3 and a plastic moderator is used to slow the neutrons. This creates an alpha particle inside the detector, and thus, neutrons can be counted.

It is worth mentioning that the ‘Geiger counter’ is commonly used to mean a handheld survey-type meter; however, the Geiger principle is in wide use by installing it to receive ‘gamma’ alarms in an area for personnel protection, and in process measurement and interlock applications. A Geiger tube is still the sensing device, but the processing electronics will have a higher degree of sophistication and reliability than that used in a handheld survey meter.

11.4.2 SCINTILLATION COUNTER

The modern electronic scintillation counter was invented in 1944 by Sir Samuel Curran (Curran, 1949). A *scintillation counter* is an instrument for detecting and measuring ionizing radiation by using the excitation effect of incident radiation on a scintillating material and detecting the resultant light pulses.

It consists of a scintillator which generates photons in response to incident radiation, a sensitive photodetector associated with a usual photomultiplier tube (PMT), a charge-coupled device (CCD) camera, or a photodiode that converts light to an electrical signal and electronics to process this signal. Scintillation counters are widely used in radiation protection, in the assay of radioactive materials, and in physics research because of its inexpensively good quantum of efficiency that can measure both the intensity and the energy of incident radiation.

11.4.2.1 Operational Mechanism of the Scintillation Counter

Whenever an ionizing particle passes into the scintillator material, atoms are excited along a track. It is observed that charged particles have track that corresponds to the path of the particle itself. It is interesting to note that the energy of gamma rays (uncharged) is converted to an energetic electron via either the photoelectric effect, Compton scattering, or pair (positron) production.

The chemistry of atomic de-excitation in the scintillator produces a multitude of low-energy photons, typically near the blue end of the visible spectrum. The quantity is proportional to the energy deposited by the ionizing particle. These can be directed to the photocathode of a photomultiplier tube, which emits, at most, one electron for each arriving photon due to the photoelectric effect. This group of primary electrons is electrostatically accelerated and focused by an electrical potential so that they strike the first dynode of the tube. The impact of a single electron on the dynode releases a number of secondary electrons, which are, in turn, accelerated to strike the second dynode. Each subsequent dynode impact releases further electrons, and so there is a current amplifying effect at each dynode stage. Each stage is at a higher potential than the previous to provide the accelerating field.

The resultant output signal at the anode is a measurable pulse for each group of photons from an original ionizing event in the scintillator that arrived at the photocathode and carries information about the energy of the original incident radiation. When it is fed to a charge amplifier which integrates the energy information, an output pulse is obtained which is proportional to the energy of the particle exciting the scintillator. The number of such pulses per unit time also gives information about the intensity of the radiation. In some applications, individual pulses are not counted, but rather, only the average current at the anode is used as a measure of radiation intensity.

The scintillator must be shielded with a thin opaque foil, such as aluminized mylar, from all ambient light so that external photons do not swamp the ionization events caused by incident radiation. It may be noted that low enough mass of shielding materials is meant to minimize undue attenuation of the incident radiation being measured. The scintillator consists of a transparent crystal, usually a phosphor, a plastic having anthracene, or an organic liquid that fluoresces when struck by ionizing radiation. Caesium iodide (CsI) in crystalline form is used as the scintillator for the detection of protons and alpha particles. Sodium iodide (NaI) containing a small amount of thallium is used as a scintillator for the detection of gamma waves, and zinc sulphide (ZnS) is widely used as a detector of alpha particles. Zinc sulphide is the material Rutherford used to perform his scattering experiment. Lithium iodide (LiI) is used in neutron detectors.

11.4.2.2 Efficiencies of the Scintillation Counter

The quantum efficiency of a gamma ray detector (per unit volume) depends upon the density of electrons in the detector, and certain scintillating materials, such as sodium iodide and bismuth germanate, achieve high electron densities as a result of the high atomic numbers of some of the elements of which they are composed. However, detectors based on semiconductors, notably hyper-pure germanium, have better intrinsic energy resolution than scintillators and are preferred where feasible for gamma ray spectrometry. In the case of neutron detectors, high efficiency is gained through the use of scintillating materials rich in hydrogen that scatter neutrons efficiently. Liquid scintillation counters are an efficient and practical means of quantifying beta radiation.

11.4.2.3 Effective Use of the Scintillation Counter

Scintillation counters are used to measure radiation in a variety of applications, including handheld radiation survey meters, personnel and environmental monitoring for radioactive contamination, medical imaging, radiometric assay, nuclear security, and nuclear plant safety.

Several products have been introduced in the market utilizing scintillation counters for the detection of potentially dangerous gamma-emitting materials during transport. These include scintillation counters designed for freight terminals, border security, ports, weigh bridge applications, scrap

metal yards, and contamination monitoring of nuclear waste. There are variants of scintillation counters mounted on pickup trucks and helicopters for rapid response in case of a security situation due to dirty bombs or radioactive waste. Handheld units are also commonly used.

11.5 TYPES OF RADIOELEMENTS AND THEIR DETECTABILITY

Naturally or artificially, radiation is reported from everywhere. Natural radioisotopes are found since, when the Earth was formed approximately 5 billion years ago, matter consisted of stable and unstable atoms. Since then, the majority of unstable atoms have disintegrated by radioactivity, and most have ended up achieving stability. However, there are still some naturally radioactive atoms:

- Radioisotopes characterized by a very long half-life, such as uranium-238 (4.5 billion years) and potassium (K)-40 (1.3 billion years). These have not had enough time to disintegrate completely since they were created.
- The radioactive descendants of the preceding elements, such as radium (Ra)-226, which is constantly being regenerated after the disintegration of uranium (U)-238. Ra-226 transforms slowly into a gas, radon (Rn)-222, which is itself radioactive.
- The radioisotopes created by the action of cosmic radiation on certain atomic nuclei. This is the case with carbon (C)-14, for example, which is constantly being formed in the atmosphere. It is known to us that cosmic radiation continuously reaches us from the universe; sometimes it is very strong.

These natural radioisotopes are present throughout the planet, in the atmosphere (carbon-14; radon-222), in the Earth's crust (uranium-238, uranium-235, radium-226, etc.), and in our food (potassium-40). That is why everything around us is radioactive. Since the dawn of time, the Earth and living beings have therefore been bathed in radioactivity. It was only recently (slightly more than 100 years ago), through the work of Henri Becquerel, that humans discovered they had always been living in this environment.

Artificial radioisotopes are produced using a cyclotron or nuclear reactor, for many different applications. Some radioisotopes (cobalt [Co]-60, iridium [Ir]-192, etc.) can be used as sources of radiation for gamma radiography or sources of irradiation for radiotherapy or industrial applications. The use of these sources is widespread in medicine and industry. Other artificial radioisotopes are created in nuclear reactors (strontium [Sr]-90, caesium [Cs]-137, etc.). Some are not used by man. They constitute what is known as nuclear waste. Being highly radioactive, these radioisotopes have to be stored under high surveillance and kept isolated from humans.

11.6 APPLICATIONS OF RADIOACTIVITY SURVEYS IN GEOPHYSICAL EXPLORATION

Radioactivity measurements are predominantly done in borehole geophysics, while radioactivity field surveys are relatively unimportant compared with other geophysical methods. Apart from uranium prospecting, radioactivity measurements may serve as an aid in geological mapping for clarifying lithostratigraphical and structural conditions. In the mining process, nuclear techniques are mainly used for recovery of uranium, coal, and oil. In uranium mining, the radioactivity of the rock is used to separate ore and waste rock and to determine the grade of the ore. In coal mining, it is important to know the thickness of the coal layer in tunnels to avoid mining waste rock.

11.6.1 GEOPHYSICAL EXPLORATION OF MINERALS AND METALS (SURFACE)

The technique of measurement of natural radiation in the Earth's surface is called the 'radiometric method' (also known as *gamma ray spectrometry*). The radiometric surveys detect and map natural radioactive emanations (γ ray) from rocks and soils. The gamma radiation takes place from

the natural decay of elements like U, Th, and K. The radiometric method is capable of detecting these elements at the surface of the ground. The common radioactive minerals are uraninite (^{238}U), monazite, thorianite (^{232}Th), rubidium (^{87}Rb) in granite-pegmatite, feldspar (^{40}K), muscovite, sylvite in acid igneous rocks, and radiocarbon (^{14}C). Exploration for these minerals by radiometric survey became important because of the demand for nuclear fuels and also for detection of associated non-radioactive deposits, such as titanium and zircon. Normally, ground radiometric surveys are used as follow-up after airborne surveys. Airborne spectrometers can use much larger crystals, measure more channels, and correct both for drift and background radiation. Ground surveys can then help isolate the sources of the airborne anomalies. The interpretation of ground survey results is mainly qualitative. This is partly due to the extremely small depth of penetration possible with the method. It is also the result of the inherent complexity of the γ ray spectra.

Two uranium-rich glacial erratics were found by Imatran Voima Oy (IVO) in Alho, north-west of Porvoo, in 1956–1957. During a car-borne radiometric survey, IVO engineers found a radioactive boulder known as IVO-9 (Äikäs, 2003) close to the road between the village of Myrskylä and the town of Porvoo. The uranium content of boulder IVO-9 is 36% U_3O_8 , and the main minerals are uraninite, calcite, quartz, hematite, chlorite, and accessory pyrite (Appelqvist and Kinnunen, 1977). Uraninite is botryoidal and usually surrounded by carbonate. Fractures in the quartz vein are partly filled by uraninite, whereas cracks in the uraninite are filled with carbonate or hematite (Äikäs, 2003). Both uraninite and hematite occur as grains and as fracture fillings. The mineral with metallic reflection is uraninite.

The terrain around boulder IVO-9 was studied carefully by IVO within a radius of a couple of hundreds of metres with radiometric ground measurements at 1 m spacing (Äikäs, 2003). In 1957, this work resulted in the discovery of a second, smaller boulder, called boulder IVO-99. The uranium content of boulder IVO-99 is 12% U_3O_8 , and its mineralogy is similar to that of boulder IVO-9. The presence of uraninite in the Alho boulders indicates hydrothermal uranium mineralization related to fault and shear zones in the bedrock. Uraninite dating gives a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age 456 Ma (A664A) from boulder IVO-9 (Vaasjoki et al., 2004). Despite the field studies by IVO in the 1950s–1960s and by the GTK in the 1970s–1980s, the bedrock source of the Alho boulders has not been discovered.

Kestell (2000), as part of an MSc in association with the BGS, found a good correlation between radiometric anomalies identified by the High Resolution Airborne Resource and Environmental Survey Phase 1 (HiRES-1) survey and colliery spoil tips (Emery et al., 2006). Thorium was found to be most elevated above background, and the presence of naturally occurring radionuclides was attributed to the dominance of shale in the spoil. The higher-resolution ground survey displays a higher degree of detail, relative to the HiRES-1 survey. Patterns of radionuclide activity displayed in the ground survey were interpreted as weathering of the plateau region, resulting in the deposition of radionuclides on the slopes of the spoil heap. Total annual effective whole-body equivalent dose rate at the site was calculated as $0.80 \pm 0.03 \text{ mSv yr}^{-1}$. Kestell concluded that this dose level could conceivably significantly increase dose to members of the public who were regularly in close contact with the area (Kestell, 2000).

Surface gamma activity superimposed over the geological map of the Beldih, South Purulia Shear Zone (SPSZ), India (Sharma et al., 2014), clearly depicts the coincidence of conductive zones and high-radioactivity zones (Sharma et al., 2014). This coincidence indicates the existence of alteration zones in this area. These zones may contain radioactive mineralization as they are also correlating with the radioactive bands identified by Katti et al. (2010) and with the borehole data. All these identified mineralized zones are almost parallel and have a collective strike oriented along the E–W direction, which is in agreement with the strike of SPSZ.

11.6.2 ENGINEERING GEOPHYSICS FOR DECIPHERING BURIED CHANNEL AND FLOW OF UNDERGROUND FLUIDS

Deep-reaching fracture zones may be special targets, especially for radon gas measurements (Figure 11.5). Radon gas emanometry has been performed since more than 60 years and has, in recent times, gained increased interest in environmental geology and geophysics. Radon originates from the uranium and thorium decay in rocks in the Earth's crust and migrates through joints and fissures of

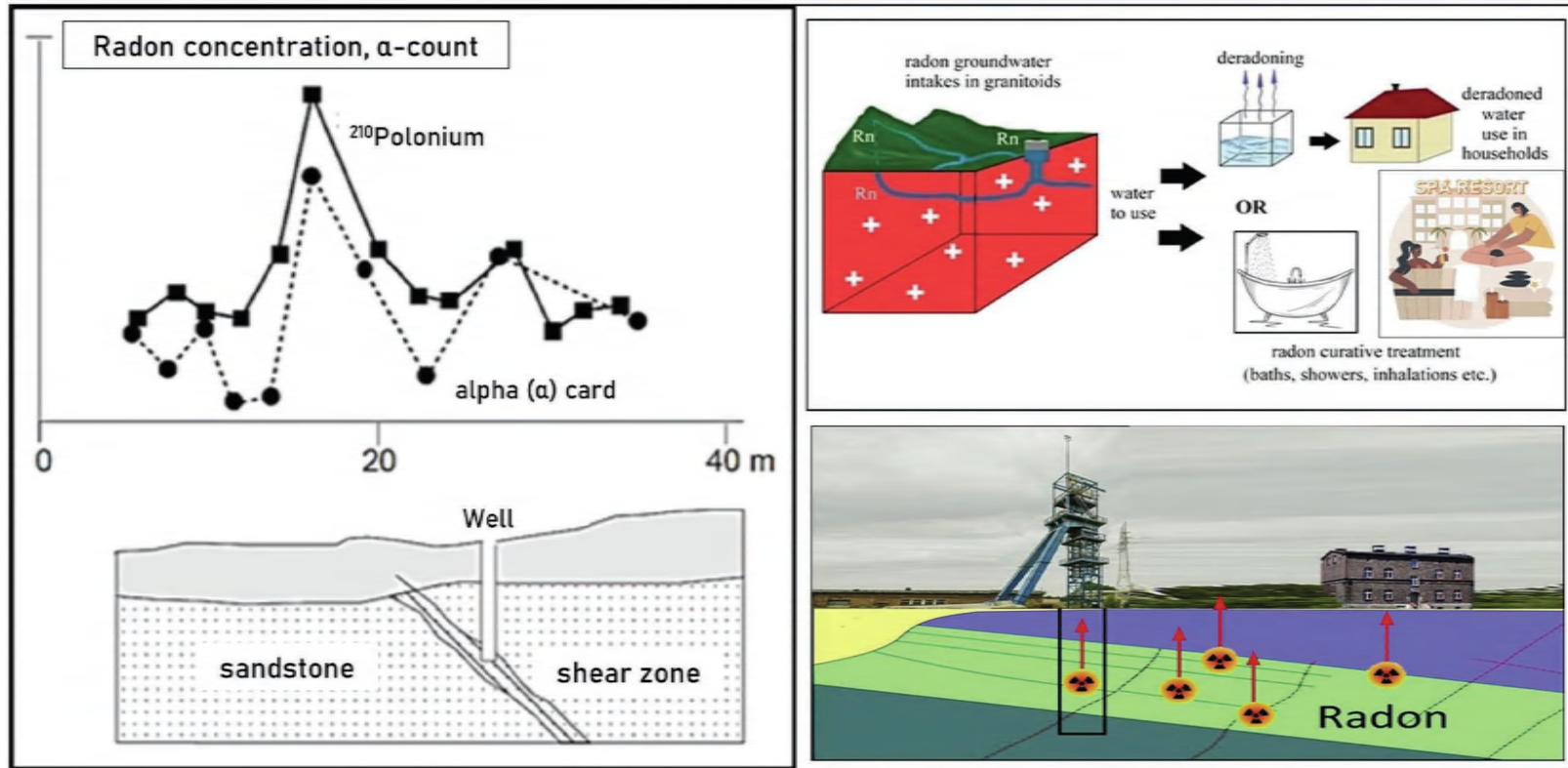


FIGURE 11.5 Exploration of groundwater associated with a fracture zone by radon measurements. Both solid-trace detector (alpha card) measurements of radon in the soil gas and α counts of the long-lived daughter element $^{210}\text{polonium}$ exhibit a distinct concentration maximum over the shear zone.

fracture zones. Groundwater migration and carrier gases are assumed to facilitate the transport, which is important because of the rapid radioactive decay of radon (3.8-day half-life). Finally, the migrated radon intermixes with the air in the soil, where the concentration is measure by the alpha radiation. The radon distribution in the soil layers is complex and not completely understood. Important factors are the lithology of the soil-forming rocks, tectonic features, the geochemistry of the radioactive elements, soil physics and soil chemistry, changing groundwater table, and the influence of topography and weather (Liang, 1986; Gregg and Holmes, 1990; Nielson et al., 1990).

11.6.3 USE OF RADIOACTIVE MEASUREMENTS AS EARTHQUAKE PRECURSORS

During the past two decades, efforts have been made to measure anomalous emanations of geo-gases in earthquake-prone regions of the world, in particular helium, radon, hydrogen, and carbon dioxide. Among them, radon has been the most preferred as earthquake precursor, because it is easily detectable.

Radon is found in nature in three different isotopes: ^{222}Rn , member of the ^{238}U series, with a half-life of 3.8 days; ^{220}Rn (also called thoron), member of the ^{232}Th series, with a half-life of 54.5 s; and ^{219}Rn , member of the ^{235}U series, with a half-life of 3.92 s. Owing to its longer half-life, the most important of them is ^{222}Rn , produced by ^{226}Ra decaying. After its production in soil or rocks, ^{222}Rn can leave the ground crust either by molecular diffusion or by convection and enters the atmosphere, where his behaviour and distribution are mainly governed by meteorological processes.

The radon decay products are radioactive isotopes of Po, Bi, Pb, and Tl, and they are easily attached to aerosol particles present in the air. In Table 11.2 are shown the principal decay

TABLE 11.2

Principal Decay Characteristics of ^{222}Rn and ^{220}Rn

Radionuclide	Half-Life	Radiation	E_{α} (MeV)	E_{γ} (MeV)
^{226}Ra	1,600 y	α	4.78 (94.3%) 4.69 (5.7%)	0.186 (83,3%)
^{222}Rn	3.824 d	α	5.49 (100%)	
^{218}Po	3.05 m	α	6.00 (100%)	
^{214}Pb	26.8 m	β, γ		0.295 (19%) 0.352 (36%)
^{214}Bi	19.7 m	β, γ		0.609 (47%) 1.120 (15%) 1,760 (16%)
^{241}Po	164 μs	α	7.69 (100%)	
^{224}Ra	3.66 d	α	5.45 (6%) 5.68 (94%)	0.241 (3.9%)
^{220}Rn	55 s	α	6.29 (100%)	
^{216}Po	0.15 s	α	6.78 (100%)	
^{212}Pb	10.64 h	β, γ		0.239 (47%) 0.300 (3.2%)
^{212}Bi	1.01 h	α, β, γ	6.05 (25%) 6.09 (10%)	0.727 (11.8%) 1.620 (2.8%)
^{212}Po	298 ns	α	8.78 (100%)	
^{208}Tl	3.05 m	β, γ		0.511 (23%) 0.583 (86%) 0.860 (12%) 2.614 (100%)

characteristics of ^{222}Rn and ^{220}Rn , including the properties of their respective parent radionuclides and their short-lived decay products (Spitsyn, 1926).

The release of radon from natural minerals has been known since the 1920s (Spitsyn, 1926), but its monitoring has more recently been used as a possible tool for earthquake prediction, because the distribution of soil gas radon concentration is closely related to the geological structure, fracture, nature of rocks, and distribution of sources. Therefore, surveying of radon concentration can prospect earthquake forecast.

In the last ten years, systematic studies on radon as precursor of geophysical events have been carried out on Mt Etna since 2001 (Immè et al., 2005; Immè et al., 2006a, Immè et al., 2006b; La Delfa et al., 2007; La Delfa et al., 2008; Morelli et al., 2006, Morelli et al., 2011). In particular, two sites were investigated among the cropping-up structural discontinuities which lie along the NE–SW direction through the volcano. One site (Biancavilla) is in the SW flank, while the other one (Vena) is in the NE flank (circles in Figure 11.6). Continuous monitoring was performed by using active systems with time resolution of 10 min. Capillary probes inserted into the soil at 1 m depth allowed to reduce influence from the meteorological parameters that were measured too. A radon anomaly was recorded before the 3 November event (M3.5), with epicentral zone close (less than 1 km) to the Vena station (NE station), also associated to evident soil fractures (Immè et al., 2005).

In Bhatsadam, Maharashtra, India, major earthquakes occurred during August 1983–July 1984. In that region, radon concentration was measured by Rastogi et al. (1986). They found an increase in radon concentration during March–April 1984, when seismicity was high enough. Precursory

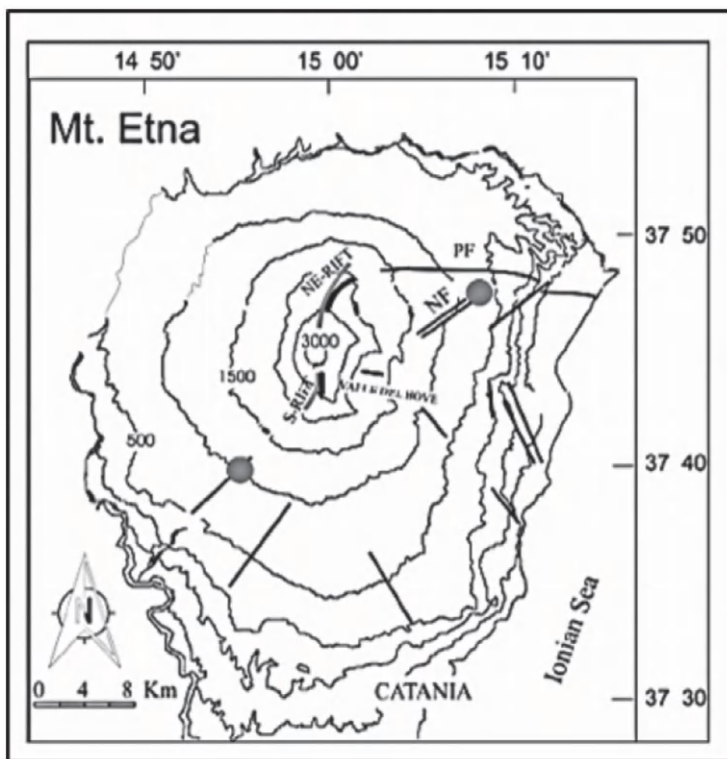


FIGURE 11.6 Mt Etna map. Circles indicate the sites where devices for continuous soil gas radon monitoring were positioned.

Source: Immè et al. (2005).

phenomena of radon in earthquake sequence were observed by Rastogi et al. (1986) and by other groups at the Osmansagar reservoir, Hyderabad, India, during January–February 1982 (Rastogi et al., 1986). An earthquake with a magnitude of 3.5 occurred on 14 January 1982, with subsequent seismic events. There was an increase of radon concentration in soil gas during February due to those high seismic activities. Singh et al. (1991) performed a daily radon monitoring in soil gas in Amritsar from 1984 to 1987. They recorded radon anomalies before different earthquakes: June 1988 (M6.8); 26 April 1986 (M5.7); July 1986 (M3.8); Kangra earthquake March 1987 (M7.0) and May 1987 (M5.0). Virk and Singh (1994) carried out daily measurements of radon in soil gas and groundwater at Palampur since 1989, and radon anomaly was recorded simultaneously in both soil gas and groundwater. Weekly integrated data also showed abnormal radon behaviour during first week of October 1991 at different recording stations. These recorded anomalies were correlated with an earthquake of magnitude 6.5 that occurred in the Uttarakashi area in October 1991.

11.7 RADIOACTIVE DATING

Certain natural radioactive elements can be used as chronometers for looking back in time. Methods of dating have been developed based on the decay of the radioactivity contained in the objects or remains being studied. It is possible to go back tens of thousands of years using carbon-14, and even further using other methods, such as thermoluminescence and uranium/thorium dating. Carbon-14 dating is used to study the history of man and his environment during a period between 5,000 and 50,000 years ago. Carbon is very widespread in our environment and, in particular, is a constituent of the carbon dioxide gas molecule present in the atmosphere. This carbon consists mainly of carbon-12. However, a small proportion of radioactive carbon-14 is naturally present. There is the same ratio of carbon-14 to carbon-12 in the atmosphere and in the living world (plants and animals) during the life of each individual because of necessary exchanges between them (respiration, photosynthesis, and food). When an organism dies, carbon-14 is no longer renewed through exchanges with the outside world. As it gradually disintegrates, the proportion contained within the organism diminishes. So by measuring the ‘carbon-14 to carbon-12’ ratio, we can ascertain the date of death. The less carbon-14 there is in the fossil to be dated, the longer ago death occurred. Radioactivity is used to date historic and prehistoric remains.

REFERENCES

- Äikäs, O., 2003. Uranium in the Itä-Uusimaa area, Finland. Geological Survey of Finland, unpublished report CM60/2003/2, p. 34.
- Appelqvist, H. and Kinnunen, K., 1977. Raportti geologisen tutkimuslaitoksen uraanitutkimuksista Askolan Porvoon alueella 1976–1977. Geological Survey of Finland, unpublished report M19/3022/77/1/10. p. 12. (in Finnish).
- Becquerel, A.H., 1896a. On the invisible rays emitted by phosphorescent bodies. *Comptes Rendus*, vol. 122, pp. 501–503.
- Becquerel, A.H., 1896b. On the rays emitted by phosphorescence. *Comptes Rendus*, vol. 122, pp. 420–421.
- CEA, 2005. *Commissariat à l'Énergie Atomique et aux Energies Alternatives*, 2005. Gif-sur-Yvette cedex, France: Communication Division, Bâtiment Siège – 91191., ISSN 1637-5408. Available at www.cea.fr.
- Curran, S.C., 1949. *Counting Tubes: Theory and Applications*. 1st edn. New York: Academic Press, vol. xi, p. 235.
- Emery, C., Davis, J.R., Hodgkinson, E. and Jones, D.G., 2006. Airborne and ground-based radiometric investigation of colliery spoil near Shirebrook, the English Midlands. British Geological Survey Internal Report, IR/05/059. p. 60.
- Glenn, F.K., 2000. *Radiation Detection and Measurement*. 3rd edn. Hoboken, NJ: John Wiley and Sons, ISBN 0471073385.
- Gregg, L.T. and Holmes, J.J., 1990. Radon detection and measurement in soil and groundwater. In: Ward, S.H. (ed), *Geotechnical and Environmental Geophysics*, vol. 1. Tulsa, OK: Society of Exploration Geophysicists, pp. 251–262.

- Immè, G., La Delfa, S., Lo Nigro, S., Morelli, D. and Patanè, G., 2005. Gas Radon emission related to geodynamic activity of Mt. Etna. *Annals of Geophysics*, vol. 48, no. 1, pp. 65–77.
- Immè, G., La Delfa, S., Lo Nigro, S., Morelli, D. and Patanè, G., 2006a. Soil Radon concentration and volcanic activity of Mt. Etna before and after the 2002 eruption. *Radiation Measurements*, vol. 41, pp. 241–245.
- Immè, G., La Delfa, S., Lo Nigro, S., Morelli, D. and Patanè, G., 2006b. Soil Radon monitoring in NE flank of Mt. Etna (Sicily). *Applied Radiation and Isotopes*, vol. 64, pp. 624–629.
- Katti, V.J., Sen, J. and Bhatt, A.K., 2010. Uranium potentiality of South Purulia Shear Zone in Eastern Indian Shield. Proceedings of the Technical Meeting on Low Grade Uranium Ore, International Atomic Energy Agency, Vienna, Austria, pp. 29–31.
- Kestell, D.J., 2000. An investigation of the distribution and radiological Impact of naturally radionuclides in industrial spoil heaps. MSc Thesis, University of Surrey. British Geological survey.
- La Delfa, S., Agostino, I., Morelli, D. and Patanè, G., 2008. Soil Radon concentration and effective stress variation at Mt Etna (Sicily) in the period January 2003–April 2005. *Radiation Measurements*, vol. 43, pp. 1299–1304.
- La Delfa, S., Immè, G., Lo Nigro, S., Morelli, D., Patanè, G. and Vizzini, F., 2007. Radon measurements in the SE and NE flank of Mt. Etna (Italy). *Radiation Measurements*, vol. 42, pp. 1404–1408.
- Liang, J.H., 1986. *Nuclear techniques in prospecting for aquifers and its effectiveness: In Contributions to the Exploration Geophysics and Geochemistry*, vol. 4. Beijing: Geol. Publishing House, (in Chinese).
- Morelli, D., Immè, G., Altamore, I., Cammisa, S., Giammanco, S., La Delfa, S., Mangano, G., Neri, M. and Patanè, G., 2011. Radionuclide measurements, via different methodologies, as tool for geophysical studies on Mt. Etna. Nuclear Instruments and Methods in Physics Research A, DOI: NIMA10.1016/j.nima.2011.01.172.
- Morelli, D., Immè, G., La Delfa, S., Lo Nigro, S. and Patanè, G., 2006. Evidence of soil Radon as tracer of magma uprising at Mt. Etna. *Radiation Measurements*, vol. 41, pp. 721–725.
- Nielson, D.L., Linpei, C. and Ward, S.H., 1990. Gamma-ray spectrometry and radon emanometry in environmental geophysics. In: Ward, S.H. (ed), *Geotechnical and Environmental Geophysics*, vol 1. Tulsa, OK: Society of Exploration Geophysicists, pp. 219–250.
- Rastogi, B.K., Chadha, R.K. and Raju, I.P., 1986. Seismicity near Bhatsa reservoir, Maharashtra, India. *Physics of the Earth and Planetary Interiors*, vol. 44, no. 2, pp. 179–199.
- Rutherford, E. and H. Geiger, H., 1908. An electrical method of counting the number of α particles from radioactive substances. *Proceedings of the Royal Society (London)*, Series A, vol. 81, no. 546, pp. 141–161.
- Sharma, S.P., Biswas, A. and Sengupta, D., 2014. Correlation of VLF-EM Data with radiometric measurements: Implications for uranium exploration around Beldih, South Purulia Shear Zone, India. *International Journal of Geophysics*, vol. 2014, p. 13, Hindawi Publishing Corporation, Article ID 969462, DOI: 10.1155/2014/969462.
- Singh, M., Ramola, R.C., Singh, B., Singh, S. and Virk, H.S., 1991. Radon anomalies: Correlation with seismic activities in northern India. In: Furlan, G. Tommasino, L. (eds), *Proceedings of the Second Workshop on Radon Monitoring in Radioprotection, Environmental and/or Earth Science*. Semantic Scholar, European Radon Program, Singapore: World Scientific, pp. 354–375.
- Spitsyn, V.I., 1926. Collection of Studies on Radium and Radioactive Ores, p. 264.
- Townsend, S.J., 1901. The conductivity produced in gases by the motion of negatively charged ions. *Philosophical Magazine*, Series 6, vol. 1, no. 2, pp. 198–227.
- Vaasjoki, M., Sundblad, K., Alm, E. and Huhma, H., 2004. Paleozoic supergene processes in the precambrian fennoscandian shield: Radiogenic isotope evidence. In: Mertanen, S. (ed), Proceedings of the 5th Nordic Paleomagnetic Workshop: Supercontinents, Remagnetizations and Geomagnetic Modeling Extended Abstract Volume. Geological Survey of Finland Report Q29.1/2004/1, pp. 129–132.
- Virk, H.S. and Singh, B., 1994. Radon recording of Uttarkashi earthquakes. *Geophysical Research Letters*, vol. 21, no.8, pp. 737–740.
- Weisstein, E.W., 2014. Radiation. Eric Weisstein's World of Physics. Champaign, IL: Wolfram Research, January.

12 Geophysical Well Logging or Borehole Logging

A *log* is defined as ‘a record of sequential data’. Geophysical well logging, also referred to as borehole geophysics, involves all techniques of lowering sensing devices in a borehole and recording the related physical properties from the boreholes by the logger stationed on the surface. In simplicity, *well logging*, also known as borehole logging, is the practice of making a detailed record (a well log) of the geologic formations in which rocks are pierced by drilling boreholes. The definition of the word *log* may be based either on the visual inspection of rock samples brought to the surface (geological logs) or on physical measurements made by instruments lowered into the hole (geophysical logs). Some types of geophysical well logs can be done during any phase of a well’s history: drilling, completing, producing, or abandoning. Geophysical well logging provides continuous objective records with values that are consistent from well to well and from time to time once the equipment is properly calibrated and standardized. Logging techniques also allow time-lapse measurements to observe changes in a dynamic system. Changes in both fluid and rock characteristics and well construction caused by pumping or injection can be determined by periodic logging. It is found that radiation logs under some conditions, like acoustic logs, are unique in providing data on aquifers or inside the coalbed through the casing.

Well logging is performed in boreholes drilled for oil and gas, groundwater, mineral, and geothermal exploration, as well as part of environmental and geotechnical studies. Geophysical well logging helps estimate some of the pertinent parameters that will be interpreted in terms of the characteristics of the rocks, the fluids contained in the rocks, as well as the construction of the well. Geophysical logs can be reinterpreted in a postmortem investigation of some geologic, hydrologic, or petrologic factors that were not considered while the hole was being drilled. The necessity of ‘log’ data is based on the fact that exploratory drill holes or wells are the only means of direct access to the surface. As such, drilling is an expensive affair to access the lithosphere, and for sampling of the rocks and fluids penetrated. Geophysical well logging is the only way to seek information in detail derived from these holes. Valid well logging also allows us to undertake the vertical and horizontal extrapolation of data derived from drill holes.

It is well documented that geophysical well logs are interpreted to determine the lithology, geometry, resistivity, formation factor, bulk density, porosity, permeability, moisture content, and specific yield of fluid-bearing rocks, which, in turn, define the source, movement, and chemical and physical characteristics of groundwater, mineral deposits, geothermal resources, and oil- and gas-bearing zones.

There are innumerable examples of logs that are used to illustrate applications and interpretation in various geoenvironmental conditions. The interpretations between various types of geophysical logging techniques with their corresponding applications are shown in Table 12.1. The graphic presentation of geophysical logs allows rapid visual interpretation and comparison at the well site. Decisions on where to set the screen and on testing procedures can be made immediately, rather than after time-consuming sample studies or laboratory analyses.

The geophysical logging technique is also associated with some of the important limitations, of which the single most important factor that has limited the use of geophysical logging in groundwater hydrology and oil-and-gas exploration is the cost in relation to the unit value of the product. The logging cost gets increased due to an increase in the number of types of logs run and in the number of runs per well. If logging is to be widely used in groundwater hydrology, a constant effort will have to be made to keep its cost low in relation to the cost of drilling and development. In order to

TABLE 12.1**Different Borehole Logging Techniques and Their Applications**

S. No.	Types of Borehole Logging Techniques	Applications
1	Electric, sonic, or calliper logs made in open holes. Nuclear logs are made in open or cased holes.	Lithology and stratigraphic correlation of aquifers and associated rocks
2	Calibrated sonic logs in open holes, calibrated neutron or gamma–gamma logs in open or cased holes.	Total porosity or bulk density
3	Calibrated long-normal resistivity logs.	Effective porosity or true resistivity
4	Gamma logs.	Clay or shale content
5	No direct measurement by logging. May be related to porosity, infectivity, sonic amplitude.	Permeability
6	Calliper, sonic, or borehole televiewer or television logs.	Secondary permeability—fractures, solution openings
7	Calibrated neutron logs.	Specific yield of unconfined aquifers
8	Possible relation to formation factor derived from electric logs.	Grain size
9	Electric, temperature of fluid conductivity in open hole or inside casing. Neutron or gamma–gamma logs in open hole or outside casing.	Location of water level or saturated zones
10	Calibrated neutron logs.	Moisture content
11	Time interval neutron logs under special circumstances or radioactive tracers.	Infiltration
12	Single-well tracer techniques—point dilution and single-well pulse. Multiwell tracer techniques.	Direction, velocity, and path of groundwater flow
13	Fluid conductivity and temperature logs, gamma logs for some radioactive wastes, fluid sampler.	Dispersion, dilution, and movement of waste
14	Injectivity profile. Flowmeter or tracer logging during pumping or injection. Temperature logs.	Source and movement of water in a well
15	Calibrated fluid conductivity and temperature in the well. Neutron chloride logging outside casing. Multielectrode resistivity.	Chemical and physical characteristics of water, including salinity, temperature, density, and viscosity
16	Gamma–gamma, calliper, collar, and perforation locator, borehole television.	Determining construction of existing wells, diameter and position of casing, perforations, and screens
17	All logs providing data on the lithology, water-bearing characteristics, and correlation and thickness of aquifers.	Guide to screen setting
18	Calliper, temperature, gamma–gamma. Acoustic for cement bond.	Cementing
19	Under some conditions, calliper, or collar locator.	Casing corrosion
20	Tracer and flowmeter.	Casing leaks and/or plugged screen

improve the cost–benefit ratio of logging, interpretation must be provided along with logs. In addition to that, one of the fundamental problems in the applications of geophysical logs is that the interpretation of many logs is more of an art than a science. The numerous environmental factors causing log response are difficult to analyze quantitatively. Even when theoretically derived equations are available, empirical data are needed to determine the unknowns in the equation. In this context, it is realized that the direct empirical methods may be more reliable because empirical methods are based on the relationship of log response to samples, core, models, or experimentally derived equations, since most geophysical logs do not have a unique response in the sense that a certain recorded deflection may be due to one lithology in one area and due to another lithology elsewhere. Because

of this constraint, background information and experience in each geologic environment are essential to guide log interpretation. Interpretation must also be based on sound knowledge of the theory of each logging technique.

12.1 WIRELINE LOGGING

The oil and gas industry uses wireline logging to obtain a continuous record of a formation's rock properties (wireline log consisting of calliper, density, and resistivity logs, adapted from US Department of the Interior, US Geological Survey, results of coalbed methane drilling in Panola County, Texas, open file report 2005–1046: geophysical logs from borehole). Wireline logging can be defined as 'the acquisition and analysis of geophysical data performed as a function of wellbore depth, together with the provision of related services'. Note that 'wireline logging' and 'mud logging' are not the same yet are closely linked through the integration of the datasets. The measurements are made in reference to TAH—true along hole depth. These and the associated analysis can then be used to infer further properties, such as hydrocarbon saturation and formation pressure, and to make further drilling and production decisions. Wireline logging is performed by lowering a 'logging tool'—or a string of one or more instruments—on the end of a wireline into an oil well (or borehole) and recording the petrophysical properties using a variety of sensors. Logging tools developed over the years measure the natural gamma ray, electrical, acoustic, stimulated radioactive responses, electromagnetic, nuclear magnetic resonance, pressure, and other properties of rocks and their contained fluids. For this chapter, they are broadly broken down by the main property that they respond to.

The data itself is recorded either at the surface (real-time mode) or in the hole (memory mode) to an electronic data format, and then either a printed record or an electronic presentation, called a 'well log', is provided to the client, along with an electronic copy of the raw data. Well logging operations can either be performed during the drilling process, to provide real-time information about the formations being penetrated by the borehole, or once the well has reached total depth and the whole depth of the borehole can be logged.

Real-time data is recorded directly against measured cable depth. Memory data is recorded against time, and then depth data is simultaneously measured against time. The two datasets are then merged using the common time base to create an 'instrument response vs depth' log. Memory-recorded depth can also be corrected in the same way as real-time corrections are made, so there should be no difference in the attainable TAH accuracy. The measured cable depth can be derived from a number of different measurements but is usually either recorded based on a calibrated wheel counter or (more accurately) using magnetic marks which provide calibrated increments of cable length. The measurements made must then be corrected for elastic stretch and temperature (Harald, 2012). There are many types of wireline logs, and they can be categorized either by their function or by the technology that they use. 'Open hole logs' are run before the oil or gas well is lined with pipe or cased. 'Cased hole logs' are run after the well is lined with casing or production pipe (Society of Professional Well Log Analysts, 1975). Wireline logs can be divided into broad categories based on the physical properties measured.

12.1.1 HISTORY OF WELL LOGGING

Conrad and Marcel Schlumberger, who founded Schlumberger Limited in 1926, are considered the inventors of electric well logging. Conrad developed the Schlumberger array, which was a technique for prospecting for metal ore deposits, and the brothers adapted that surface technique to subsurface applications. On 5 September 1927, a crew working for Schlumberger lowered an electric sonde or tool down a well in Pechelbronn, Alsace, France, creating the first well log. In modern terms, the first log was a resistivity log that could be described as a 3.5 m upside-down lateral log (Hilchie, 1990).

In 1931, Henri George Doll and G. Dechatre, working for Schlumberger, discovered that the galvanometer wiggled even when no current was being passed through the logging cables down in the well. This led to the discovery of spontaneous potential (SP), which was as important as the ability to measure resistivity. The SP effect was produced naturally by the borehole mud at the boundaries of permeable beds. By simultaneously recording SP and resistivity, loggers could distinguish between permeable oil-bearing beds and impermeable nonproducing beds.

In 1940, Schlumberger invented the spontaneous potential dipmeter; this instrument allowed the calculation of the dip and direction of the dip of a layer. The basic dipmeter was later enhanced by the resistivity dipmeter (1947) and the continuous resistivity dipmeter (1952). Oil-based mud (OBM) was first used in Rangeley Field, Colorado, in 1948. Normal electric logs require conductive or water-based mud, but OBMs are nonconductive. The solution to this problem was the induction log, developed in the late 1940s. The introduction of the transistor and integrated circuits in the 1960s made electric logs vastly more reliable. Computerization allowed much faster log processing and dramatically expanded log data-gathering capacity. The 1970s brought more logs and computers. These included combo-type logs, where resistivity logs and porosity logs were recorded in one pass in the borehole. The two types of porosity logs (acoustic logs and nuclear logs) date originally from the 1940s. Sonic logs grew out of technology developed during World War II. Nuclear logging has supplemented acoustic logging, but acoustic or sonic logs are still run on some combination logging tools.

Nuclear logging was initially developed to measure the natural gamma radiation emitted by underground formations. However, the industry quickly moved to logs that actively bombard rocks with nuclear particles. The gamma ray log, measuring natural radioactivity, was introduced by Well Surveys Inc. in 1939, and the WSI neutron log came in 1941. The gamma ray log is particularly useful as shale beds, which often provide a relatively low-permeability cap over hydrocarbon reservoirs, usually display a higher level of gamma radiation. These logs were important because they can be used in cased wells (wells with production casing). WSI quickly became part of Lane-Wells. During World War II, the US government gave a near-wartime monopoly on open-hole logging to Schlumberger, and a monopoly on cased-hole logging to Lane-Wells. Nuclear logs continued to evolve after the war.

After the discovery of nuclear magnetic resonance by Bloch and Purcell in 1946, the nuclear magnetic resonance log using the Earth's field was developed in the early 1950s by Chevron and Schlumberger. The NMR log was a scientific success but an engineering failure. More recent engineering developments by NUMAR (a subsidiary of Halliburton) in the 1990s have resulted in continuous NMR logging technology which is now applied in the oil and gas, water, and metal exploration industries (Kleinberg and Jackson, 2001). Many modern oil and gas wells are drilled directionally. At first, loggers had to run their tools somehow attached to the drill pipe if the well was not vertical. Modern techniques now permit continuous information at the surface. This is known as logging while drilling (LWD), or measurement while drilling (MWD). MWD logs use mud pulse technology to transmit data from the tools on the bottom of the drill string to the processors at the surface.

12.1.2 ELECTRICAL LOGGING

12.1.2.1 Resistivity Log

Resistivity logging measures the subsurface electrical resistivity, which is the ability to impede the flow of electric current. This helps differentiate between formations filled with salty waters (good conductors of electricity) and those filled with hydrocarbons (poor conductors of electricity). Resistivity and porosity measurements are used to calculate water saturation. Resistivity is expressed in ohms or ohms/metre and is frequently charted on a logarithm scale versus depth because of the large range of resistivity. The distance from the borehole penetrated by the current varies with the tool, from a few centimetres to 1 m.

12.1.2.2 Borehole Image Logs

The term ‘borehole imaging’ refers to those logging and data processing methods that are used to produce centimetre-scale images of the borehole wall and the rocks that make it up. The context is, therefore, that of an open hole, but some of the tools are closely related to their cased-hole equivalents. Borehole imaging has been one of the most rapidly advancing technologies in wireline well logging. The applications range from detailed reservoir description through reservoir performance to enhanced hydrocarbon recovery. Specific applications are fracture identification (Taherdangkoo and Abdideh, 2016), analysis of small-scale sedimentological features, evaluation of net pay in thinly bedded formations, and the identification of breakouts (irregularities in the borehole wall that are aligned with the minimum horizontal stress and appear where stresses around the wellbore exceed the compressive strength of the rock). The subject area can be classified into four parts:

1. Optical imaging
2. Acoustic imaging
3. Electrical imaging
4. Methods that draw on both acoustic and electrical imaging techniques using the same logging tool

(Prensky, 1999)

12.1.3 POROSITY LOGS

Porosity logs measure the fraction or percentage of pore volume in a volume of rock. Most porosity logs use either acoustic or nuclear technology. Acoustic logs measure the characteristics of sound waves propagated through the wellbore environment. Nuclear logs utilize nuclear reactions that take place in the downhole logging instrument or in the formation. Nuclear logs include density logs and neutron logs, as well as gamma ray logs, which are used for correlation (Sengel, 1981). The basic principle behind the use of nuclear technology is that a neutron source placed near the formation whose porosity is being measured will result in neutrons being scattered by the hydrogen atoms, largely those present in the formation fluid. Since there is little difference in the neutrons scattered by hydrocarbons or water, the porosity measured gives a figure close to the true physical porosity, whereas the figure obtained from electrical resistivity measurements is that due to the conductive formation fluid. The difference between neutron porosity and electrical porosity measurements therefore indicates the presence of hydrocarbons in the formation fluid.

12.1.3.1 Density Log

Density logging is a well logging tool that can provide a continuous record of a formation’s bulk density along the length of a borehole. In geology, *bulk density* is a function of the density of the minerals forming a rock (i.e. matrix) and the fluid enclosed in the pore spaces. This is one of three well logging tools that are commonly used to calculate porosity, the other two being sonic logging and neutron porosity logging.

The tool was initially developed in the 1950s and was in use throughout the hydrocarbon industry by the 1960s. A type of an active nuclear tool, a radioactive source, and a detector are lowered down the borehole, and the source emits medium-energy gamma rays into the formation. Radioactive sources are typically a directional Cs-137 source. These gamma rays interact with electrons in the formation and are scattered in an interaction known as Compton scattering. The number of scattered gamma rays that reach the detector, placed at a set distance from the emitter, is related to the formation’s electron density, which itself is related to the formation’s bulk density ρ_{bulk} via

$$\rho_e = 2 \rho_{bulk} \frac{Z}{A}$$

where Z is the atomic number and A is the molecular weight of the compound. For most elements, $-\frac{Z}{A}$ is about $\frac{1}{2}$ (except for hydrogen, where this ratio is 1). The electron density (ρ_e) in gm/cm³ determines the response of the density tool.

The tool itself initially consisted of a radioactive source and a single detector, but this configuration is susceptible to the effects of the drilling fluid. In a similar way to how the sonic logging tool was improved to compensate for borehole effects, density logging now conventionally uses two or more detectors. In a two-detector configuration, the short-spaced detector has a much shallower depth of investigation than the long-spaced detector, so it is used to measure the effect that the drilling fluid has on the gamma ray detection. This result is then used to correct the long-spaced detector.

Assuming that the measured bulk density (ρ_{bulk}) only depends on matrix density (ρ_{matrix}) and fluid density (ρ_{fluid}), and that these values are known along the wellbore, porosity (ϕ) can be inferred by the formula

$$\phi = \frac{\rho_{matrix} - \rho_{bulk}}{\rho_{matrix} + \rho_{bulk}}$$

Common values of matrix density (ρ_{matrix}) in gm/cm³ are:

Quartz sand: 2.65
Limestone: 2.71
Dolomite: 2.87

This method is the most reliable porosity indicator for sandstones and limestones because their density is well known. On the other hand, the density of clay minerals such as mudstone is highly variable, depending on depositional environment, overburden pressure, type of clay mineral, and many other factors. It can vary from 2.1 (montmorillonite) to 2.76 (chlorite), so this tool is not as useful for determining their porosity. A fluid bulk density (ρ_{fluid}) of 1 gm/cm³ is appropriate where the water is fresh, but highly saline water has a slightly higher density, and lower values should be used for hydrocarbon reservoirs, depending on the hydrocarbon density and residual saturation. In some applications, hydrocarbons are indicated by the presence of abnormally high log porosities.

12.1.3.2 Neutron Porosity

In the field of formation evaluation, porosity is one of the key measurements to quantify oil and gas reserves. The neutron porosity measurement employs a neutron source to measure the hydrogen index in a reservoir, which is directly related to porosity. The *hydrogen index* (HI) of a material is defined as the ratio of the concentration of hydrogen atoms per cm³ in the material to that of pure water at 75°F. As hydrogen atoms are present in both water- and oil-filled reservoirs, measurement of the amount allows estimation of the amount of liquid-filled porosity.

Neutrons are typically emitted by a radioactive source, such as americium beryllium (Am-Be) or plutonium beryllium (Pu-Be), or generated by electronic neutron generators, such as a minitron (Figure 12.1).

Fast neutrons are emitted by these sources with energy ranges from 4 MeV to 14 MeV and inelastically interact with matter. Once slowed down to 2 MeV, they start to scatter elastically and slow down further until the neutrons reach a thermal energy level of about 0.025 eV. When thermal neutrons are then absorbed, gamma rays are emitted. A suitable detector, positioned at a certain distance from the source, can measure either epithermal neutron population, thermal neutron population, or the gamma rays emitted after the absorption (Etnyre, 1989). Boron is known to cause anomalously low neutron tool count rates due to it having a high capture cross section for thermal neutron absorption. An increase in hydrogen concentration in clay minerals has a similar effect on

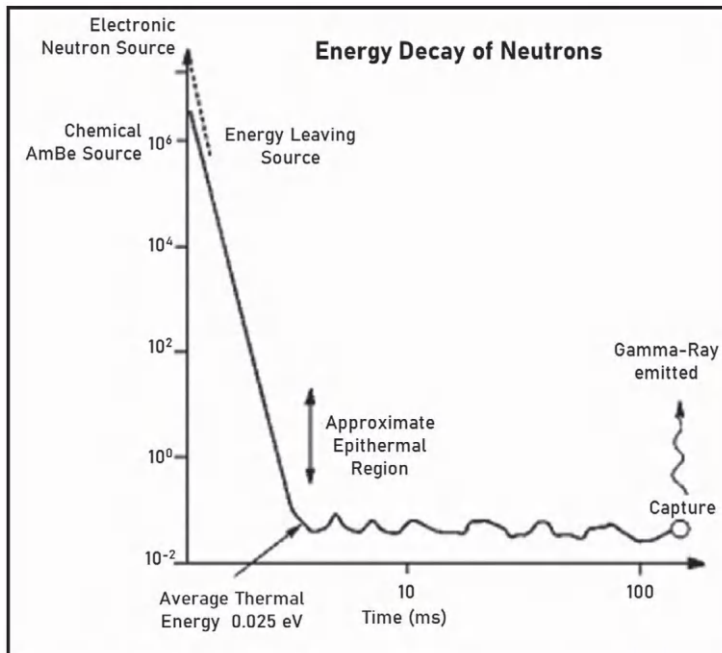


FIGURE 12.1 Energy decay of neutrons.

Source: Dewan (1986).

the count rate. A proportional counter filled with helium-3 (He-3) is the most common epithermal and thermal neutron detector. Helium has a high neutron capture cross section and produces the following reaction when interacting with a neutron:



The mechanics of elastic collisions predict that the maximum energy transfer occurs during the collisions of two particles of equal mass. Therefore, a hydrogen atom (H) will cause a neutron to slow down the most, as they are of roughly equal mass. As hydrogen is fundamentally associated to the amount of water and/or oil present in the pore space, the measurement of neutron population within the investigated volume is directly linked to porosity. Determination of porosity is one of the most important uses of neutron porosity log. Correction parameters for lithology, borehole parameters, and others are necessary for accurate porosity determination, as follows:

- Borehole size
- Borehole salinity
- Borehole temperature and pressure
- Mud cake
- Mud weight
- Formation salinity
- Tool standoff from borehole wall

Subject to various assumptions and corrections, values of apparent porosity can be derived from any neutron log. One cannot underestimate the slowdown of neutrons by other elements even if they are less-effective. Certain effects, such as lithology, clay content, and amount and type of

hydrocarbons, can be recognized and corrected for only if additional porosity information is available, for example, from sonic and/or density log. Any interpretation of a neutron log alone should be undertaken with a realization of the uncertainties involved.

The quantitative response of neutron tool to gas or light hydrocarbon depends primarily on hydrogen index and ‘excavation effect’. The hydrogen index can be estimated from the composition and density of the hydrocarbons. Given a fixed volume, gas has considerably lower hydrogen concentration. When pore spaces in the rock are excavated and replaced with gas, the formation has smaller neutron-slowing characteristic, hence the term ‘excavation effect’. If this effect is ignored, a neutron log will show a low porosity value. This characteristic allows a neutron porosity log to be used with other porosity logs (such as a density log) to detect gas zones and identify gas–liquid contacts.

12.1.3.3 Sonic Log

Sonic logging is a well logging tool that provides a formation’s interval transit time, designated as Δt , which is a measure of a formation’s capacity to transmit seismic waves. Geologically, this capacity varies with lithology and rock textures, most notably decreasing with an increasing effective porosity. This means that a sonic log can be used to calculate the porosity of a formation if the seismic velocity of the rock matrix, V_{mat} , and pore fluid, V_f , are known, which is very useful for hydrocarbon exploration. The velocity is calculated by measuring the travel time from the piezo-electric transmitter to the receiver, normally with the units microsecond per foot (a measure of slowness). To compensate for the variations in drilling mud thickness, there are actually two receivers, one near and one far. This is because the travel time within the drilling mud will be common for both, so the travel time within the formation is given by

$$\Delta t = t_{far} - t_{near}$$

where t_{far} is the travel time to the far receiver, and t_{near} the travel time to the near receiver.

If it is necessary to compensate for tool tilt and variations in the borehole width, then both up–down and down–up arrays can be used and an average can be calculated. Overall, this gives a sonic log that can be made up of one or two pulse generators and two or four detectors, all located in a single unit, called a ‘sonde’, which is lowered down the well (Sheriff and Geldart, 1995). An additional way in which the sonic log tool can be altered is by increasing or decreasing the separation between the source and the receivers. This gives deeper penetration and overcomes the problem of low-velocity zones posed by borehole wall damage. The returning signal is a wave train and not a sharp pulse, so the detectors are only activated at a certain signal threshold. Sometimes, both detectors will not be activated by the same peak (or trough), and the next peak (or trough) wave will activate one of them instead. This type of error is called cycle skipping and is easily identified because the time difference is equal to the time interval between successive pulse cycles.

Many relationships between travel time and porosity have been proposed, the most commonly accepted of which is the Wyllie time average equation. The equation basically holds that the total travel time recorded on the log is the sum of the time the sonic wave spends travelling the solid part of the rock, called the rock matrix, and the time spent travelling through the fluids in the pores. This equation is empirical and makes no allowance for the structure of the rock matrix or the connectivity of the pore spaces, so extra corrections can often be added to it. The Wyllie time average equation is (Wyllie et al., 1958):

$$\frac{1}{V} = \frac{\phi}{V_f} + \frac{1-\phi}{V_{mat}}$$

where V is the seismic velocity of the formation, V_f the seismic velocity of the pore fluid, V_{mat} the seismic velocity of the rock matrix, and ϕ the porosity. The accuracy of sonic logs is rather poor,

evident by the fact that regular- and long-spaced log measurements often conflict, and this should be taken into account when there are disagreements between seismic data and sonic log data (Sheriff and Geldart, 1995). Other considerations include that the resolution of sonic logs is on a scale of inches, whereas seismic reflection data resolution is on the scale of metres, and the two methods use significantly different frequency ranges, so travel times will vary due to dispersion. In order to investigate how the varying size of a borehole has affected a sonic log, the results can be plotted against those of a calliper log. To improve the tie between well data and seismic data, a 'check-shot' survey is often used to generate a calibrated sonic log. A geophone, or an array of geophones, is lowered down the borehole, with a seismic source located at the surface. The seismic source is fired with the geophones at a series of different depths, with the interval transit times being recorded (Schlumberger oilfield glossary). This is often done during the acquisition of a vertical seismic profile. Sonic logs are also used in mineral exploration, especially in the exploration for iron and potassium.

12.1.4 LITHOLOGY LOGS

12.1.4.1 Gamma Ray Log

Gamma ray logging is a method of measuring naturally occurring gamma radiation to characterize the rock or sediment in a borehole or drill hole. It is a wireline logging method used in mining, mineral exploration, and water well drilling; for formation evaluation in oil and gas well drilling; and for other related purposes. Different types of rock emit different amounts and different spectra of natural gamma radiation. In particular, shales usually emit more gamma rays than other sedimentary rocks, such as sandstone, gypsum, salt, coal, dolomite, or limestone, because radioactive potassium is a common component in their clay content, and because the cation exchange capacity of clay causes them to absorb uranium and thorium. This difference in radioactivity between shales and sandstones/carbonate rocks allows the gamma ray tool to distinguish between shales and non-shales (Darling, 2005). But it cannot distinguish between carbonates and sandstone, as they both have similar deflections on the gamma ray log. Thus, gamma ray logs cannot be said to make good lithological logs by themselves, but in practice, gamma ray logs are compared side by side with stratigraphic logs.

The gamma ray log, like other types of well logging, is done by lowering an instrument down the drill hole and recording gamma radiation variation with depth. In the United States, the device most commonly records measurements at $\frac{1}{2}$ foot intervals. Gamma radiation is usually recorded in American Petroleum Institute (API) units, a measurement originated by the petroleum industry. Gamma rays attenuate according to the diameter of the borehole mainly because of the properties of the fluid filling the borehole, but because gamma logs are generally used in a qualitative way, amplitude corrections are usually not necessary. Three elements and their decay chains are responsible for the radiation emitted by rock: potassium, thorium, and uranium. Shales often contain potassium as part of their clay content and tend to absorb uranium and thorium as well. A common gamma ray log records the total radiation and cannot distinguish between the radioactive elements, while a spectral gamma ray log can.

For standard gamma ray logs, the measured value of gamma ray radiation is calculated from a concentration of uranium in ppm, thorium in ppm, and potassium in weight percent: for example, $GR\ API = 8 \times \text{uranium concentration in ppm} + 4 \times \text{thorium concentration in ppm} + 16 \times \text{potassium concentration in weight percent}$. Due to the weighted nature of uranium concentration in the GR API calculation, anomalous concentrations of uranium can cause clean sand reservoirs to appear shaly. For this reason, spectral gamma ray is used to provide an individual reading for each element so that anomalous concentrations can be found and properly interpreted.

An advantage of the gamma log over some other types of well logs is that it works through the steel and cement walls of cased boreholes. Although concrete and steel absorb some of the gamma radiation, enough travels through the steel and cement to allow for qualitative determinations. In

some places, non-shales exhibit elevated levels of gamma radiation. For instance, sandstones can contain uranium minerals, potassium feldspar, clay filling, or lithic fragments that cause the rock to have higher-than-usual gamma readings. Coal and dolomite may contain absorbed uranium. Evaporite deposits may contain potassium minerals such as sylvite and carnallite. When this is the case, spectral gamma ray logging should be done to identify the source of these anomalies.

Spectral logging is the technique of measuring the spectrum, or number and energy, of gamma rays emitted via natural radioactivity of the rock formation. There are three main sources of natural radioactivity on Earth: potassium (^{40}K), thorium (principally ^{232}Th and ^{230}Th), and uranium (principally ^{238}U and ^{235}U). These radioactive isotopes each emit gamma rays that have a characteristic energy level measured in MeV. The quantity and energy of these gamma rays can be measured by a scintillometer. A log of the spectroscopic response to natural gamma ray radiation is usually presented as a total gamma ray log that plots the weight fraction of potassium (%), thorium (ppm), and uranium (ppm). The primary standards for the weight fractions are geological formations with known quantities of the three isotopes. Natural gamma ray spectroscopy logs became routinely used in the early 1970s, although they had been studied from the 1950s. The characteristic gamma ray line that is associated with each radioactive component:

- Potassium: gamma ray energy 1.46 MeV
- Thorium series: gamma ray energy 2.61 MeV
- Uranium-radium series: gamma ray energy 1.76 MeV

Another example of the use of spectral gamma ray logs is to identify specific clay types, like kaolinite or illite. This may be useful for interpreting the environment of deposition, as kaolinite can form from feldspars in tropical soils by leaching of potassium, and low potassium readings may thus indicate the presence of one or more palaeosols. The identification of specific clay minerals is also useful for calculating the effective porosity of reservoir rock. Gamma ray logs are also used in mineral exploration, especially the exploration for phosphates, uranium, and potassium salts.

12.1.4.2 Self/Spontaneous Potential Log

Spontaneous potential log, commonly called the self-potential log, or SP log, is a passive measurement taken by oil industry well loggers to characterize rock formation properties. The log works by measuring small electric potentials (measured in millivolts) between depths within the borehole and a grounded electrode at the surface (Etnyre, 1989). Conductive borehole fluids are necessary to create an SP response, so the SP log cannot be used in nonconductive drilling muds (e.g. oil-based mud) or air-filled holes (Fakhry, 2008).

The change in voltage through the wellbore is caused by a build-up of charge on the wellbore walls. Clays and shales (which are composed predominantly of clays) will generate one charge, and permeable formations such as sandstone will generate an opposite one. Spontaneous potentials occur when two aqueous solutions with different ionic concentrations are placed in contact through a porous, semipermeable membrane. In nature, ions tend to migrate from high to low ionic concentrations. In the case of SP logging, the two aqueous solutions are the wellbore fluid (drilling mud) and the formation water (connate water). The potential opposite shales are called the baseline and typically shift only slowly over the depth of the borehole.

The relative salinity of the mud and the formation water will determine the way the SP curve will deflect opposite a permeable formation. Generally, if the ionic concentration of the wellbore fluid is less than the formation fluid, then the SP reading will be more negative (usually plotted as a deflection to the left). If the formation fluid has an ionic concentration less than the wellbore fluid, the voltage deflection will be positive (usually plotted as an excursion to the right). The amplitudes of the line made by the changing SP will vary from formation to formation and will not give a definitive answer to how permeable or porous the formation it is logging. The presence of hydrocarbons (e.g. oil, natural gas, and condensate) will reduce the response on an SP log because the interstitial water contact with the wellbore

fluid is reduced. This phenomenon is called hydrocarbon suppression and can be used to diagnose rocks for commercial potential. The SP curve is usually 'flat' opposite shale formations because there is no ion exchange due to the low-permeability, low-porosity properties (tight), thus creating a baseline. Tight rocks other than shale (e.g. tight sandstones, tight carbonates) will also result in poor or no response on the SP curve because of no ion exchange. The SP tool is one of the simplest tools and is generally run as standard when logging a hole, along with the gamma ray. SP data can be used to find:

- Depths of permeable formations
- The boundaries of these formations
- Correlation of formations when compared with data from other analogue wells
- Values for the formation water resistivity

The SP curve can be influenced by various factors both in the formation and introduced into the wellbore by the drilling process. These factors can cause the SP curve to be muted or even inverted, depending on the situation.

- Formation bed thickness
- Resistivities in the formation bed and the adjacent formations
- Resistivity and make-up of the drilling mud
- Wellbore diameter
- The depth of invasion by the drilling mud into the formation

Mud invasion into the permeable formation can cause the deflections in the SP curve to be rounded off and to reduce the amplitude of thin beds (The Australian Drilling, 1997). A smaller wellbore will cause, like a mud filtrate invasion, the deflections on the SP curve to be rounded off and decrease the amplitude opposite thin beds, while a larger-diameter wellbore has the opposite effect. If the salinity of the mud filtrate is greater than that of formation water, the SP currents will flow in the opposite direction. In that case, SP deflection will be positive towards the right. Positive deflections are observed for freshwater-bearing formations.

12.1.5 MISCELLANEOUS LOGGING

12.1.5.1 Calliper Log

A *calliper log* is a well logging tool that provides a continuous measurement of the size and shape of a borehole along its depth and is commonly used in hydrocarbon exploration when drilling wells. The measurements that are recorded can be an important indicator of cave-ins or shale swelling in the borehole, which can affect the results of other well logs. The calliper tool measures the variation in borehole diameter as it is withdrawn from the bottom of the hole, using two or more articulated arms that push against the borehole wall. Each arm is typically connected to a potentiometer which causes the resistance to change as the diameter of the borehole changes, creating a varying electrical signal that represents the changing shape of the borehole. This variation in output is translated into changes of diameter after a simple calibration and the calliper log is printed as a continuous series of values of hole diameter with depth. Known challenges with calliper logging include borehole spiralling. The position of the drill bit may precess as it drills, leading to spiralling shapes in the wellbore wall, as if the hole had been drilled by a screw. If the arms of the calliper log follow the grooves of the spiral, it will report too high an average diameter. Moving in and out of the grooves, the calliper will give erratic or periodically varying readings. In most cases, the borehole's circumference will not be a perfect circle, and therefore, a calliper tool with several arms is required to obtain a true understanding of the size and shape of the borehole. The borehole can change to an oval shape after drilling, which can cause the calliper log to overestimate the size of the borehole if its arms get stuck axis.

12.1.5.2 Nuclear Magnetic Resonance Log

Nuclear magnetic resonance (NMR) logging is a type of well logging that uses the NMR response of a formation to directly determine its porosity and permeability, providing a continuous record along the length of the borehole (Gluyas and Swarbrick, 2004). NMR logging exploits the large magnetic moment of hydrogen, which is abundant in rocks in the form of water. The NMR signal amplitude is proportional to the quantity of hydrogen nuclei present in a formation and can be calibrated to give a value for porosity that is free from lithology effects. Uniquely, a petrophysicist can also analyze the rate of decay of the NMR signal amplitude to obtain information on the permeability of the formation—a crucial quantity in hydrocarbon exploration. The most important mechanism affecting NMR relaxation is grain surface relaxation. Molecules in fluids are in constant Brownian motion, diffusing about the pore space and bouncing off the grain surfaces. Upon interaction with the grain surface, hydrogen protons can transfer some nuclear spin energy to the grain (contributing to T1 relaxation) or irreversibly dephase (contributing to T2 relaxation). Therefore, the speed of relaxation most significantly depends on how often the hydrogen nuclei collide with the grain surface, and this is controlled by the surface-to-volume ratio of the pore in which the nuclei are located. Collisions are less frequent in larger pores, resulting in a slower decay of the NMR signal amplitude and allowing a petrophysicist to understand the distribution of pore sizes.

12.1.5.3 Spectral Noise Log

Spectral noise logging (SNL) is an acoustic noise measuring technique used in oil and gas wells for well integrity analysis, identification of production and injection intervals, and hydrodynamic characterization of the reservoir. (*Noise* is unwanted sound considered unpleasant, loud, or disruptive to hearing. From a physics standpoint, noise is indistinguishable from sound, as both are vibrations through a medium, such as air or water. The difference arises when the brain receives and perceives a sound. Acoustic noise is any sound in the acoustic domain, either deliberate, for example, music or speech, or unintended. In contrast, noise in electronics may not be audible to the human ear and may require instruments for detection.). An *oil well* is a boring in the Earth that is designed to bring petroleum oil hydrocarbons to the surface. Usually, some natural gas is released as associated petroleum gas along with the oil. A well that is designed to produce only gas may be termed a *gas well*. SNL records acoustic noise generated by fluid or gas flow through the reservoir or leaks in downhole well components.

Noise logging tools have been used in the petroleum industry for several decades. (The petroleum industry, also known as the oil industry, or the oil patch, includes the global processes of exploration, extraction, refining, transporting—often by oil tankers and pipelines—and marketing of petroleum products. The largest-volume products of the industry are fuel oil and gasoline, or petrol. Petroleum, which is oil, is also the raw material for many chemical products, including pharmaceuticals, solvents, fertilizers, pesticides, synthetic fragrances, and plastics. The extreme monetary value of oil and its products has led to it being known as ‘black gold’. The industry is usually divided into three major components: upstream, midstream, and downstream.) As far back as 1955, an acoustic detector was proposed for use in well integrity analysis to identify casing holes (Enright, 1955). Over many years, downhole noise logging tools proved effective in inflow and injectivity profiling of operating wells, leak detection, location of cross flows behind casing (McKinley, 1994), and even in determining reservoir fluid compositions. Robinson (1974) described how noise logging can be used to determine effective reservoir thickness.

12.2 LOGGING WHILE DRILLING

In the 1970s, a new approach to wireline logging was introduced in the form of logging while drilling (LWD). (LWD is a technique of conveying well logging tools into the well borehole downhole as part of the bottom hole assembly, or BHA.) (Barron et al., 1989). This technique provides similar

well information to conventional wireline logging, but instead of sensors being lowered into the well at the end of a wireline cable, the sensors are integrated into the drill string (a drill string on a drilling rig is a column, or string, of drill pipe that transmits drilling fluid, via the mud pumps, and torque, via the kelly drive or top drive, to the drill bit). The measurements are made in real time, whilst the well is being drilled. This allows drilling engineers and geologists to quickly obtain information, such as porosity, resistivity, hole direction, and weight-on-bit, and they can use this information to make immediate decisions about the future of the well and the direction of drilling.

In LWD, measured data is transmitted to the surface in real time via pressure pulses in the well's mud fluid column. This mud telemetry method provides a bandwidth of less than 10 bits per second, although, as drilling through rock is a fairly slow process, data compression techniques mean that this is an ample bandwidth for real-time delivery of information. A higher sample rate of data is recorded into memory and retrieved when the drill string is withdrawn at bit changes. High-definition downhole and subsurface information is available through networked or wired drill pipe that delivers memory-quality data in real time (Ali et al., 2008).

12.3 CORROSION WELL LOGGING

Throughout the life of the wells, integrity controls of the steel and cemented column (casing and tubing) are performed using callipers and thickness gauges. These advanced technical methods use nondestructive technologies such as ultrasonic, electromagnetic, and magnetic transducers (Stéphane, 2010).

12.4 MEMORY LOG

This method of data acquisition involves recording the sensor data into a downhole memory, rather than transmitting 'real time' to the surface. There are some advantages and disadvantages to this memory option.

- The tools can be conveyed into wells where the trajectory is deviated or extended beyond the reach of conventional electric wireline cables. This can involve a combination of weight-to-strength ratio of the electric cable over this extended reach. In such cases, the memory tools can be conveyed on pipe or coil tubing.
- The type of sensors is limited in comparison to those used on an electric line and tends to be focussed on the cased hole, the production stage of the well. Although there are now developed some memory 'open hole' compact formation evaluation tool combinations. These tools can be deployed and carried downhole concealed internally in drill pipe to protect them from damage while running in the hole, and then 'pumped' out the end at depth to initiate logging. Other basic open-hole formation evaluation memory tools are available for use in 'commodity' markets on slickline to reduce costs and operating time.
- In cased-hole operation, there is normally a 'slickline' intervention unit. This uses a solid mechanical wire (0.072–0.125 in in OD) to manipulate or otherwise carry out operations in the wellbore completion system. Memory operations are often carried out on this slickline conveyance in preference to mobilizing a full-service electric wireline unit.
- Since the results are not known until returned to surface, any real-time well dynamic changes cannot be monitored real time. This limits the ability to modify or change the well downhole production conditions accurately during the memory logging by changing the surface production rates. Something that is often done in electric line operations.
- Failure during recording is not known until the memory tools are retrieved. This loss of data can be a major issue on large offshore (expensive) locations. On land locations (e.g. South Texas, USA), where there is what is called a 'commodity' oil service sector, where logging often is without the rig infrastructure, this is less problematic, and logs are often run again without issue.

12.5 CORING

Coring is the process of obtaining an actual sample of a rock formation from the borehole. There are two main types of coring: ‘full coring’, in which a sample of rock is obtained using a specialized drill bit as the borehole is first penetrating the formation, and ‘sidewall coring’, in which multiple samples are obtained from the side of the borehole after it has penetrated through a formation. The main advantages of sidewall coring over full coring are that it is cheaper (drilling does not have to be stopped) and multiple samples can be easily acquired, with the main disadvantages being that there can be uncertainty in the depth at which the sample was acquired and the tool can fail to acquire the sample.

12.6 MUD LOGGING

Mud logging is the creation of a detailed record (well log) of a borehole by examining the cuttings of rock brought to the surface by the circulating drilling medium, most commonly drilling mud. (In geotechnical engineering, drilling fluid, also called drilling mud, is used to aid the drilling of boreholes into the Earth. Often used while drilling oil and natural gas wells and on exploration drilling rigs, drilling fluids are also used for much simpler boreholes, such as water wells. One of the functions of drilling mud is to carry cuttings out of the hole.) Mud logging is usually performed by a third-party mud logging company.

This provides well owners and producers with information about the lithology and fluid content of the borehole while drilling. Historically, it is the earliest type of well log. Under some circumstances, compressed air is employed as a circulating fluid, rather than mud. Although most commonly used in petroleum exploration, mud logging is also sometimes used when drilling water wells and in other mineral exploration, where drilling fluid is the circulating medium used to lift cuttings out of the hole. In hydrocarbon exploration, hydrocarbon surface gas detectors record the level of natural gas brought up in the mud. In practice, significance is placed only on relative changes in the gas concentrations detected (Bourgoyne, 1986). A mobile laboratory is situated by the mud logging company near the drilling rig or on deck of an offshore drilling rig, or on a drill ship.

12.7 DOWNHOLE AND UNDERGROUND GEOPHYSICAL METHODS FOR NICKEL SULPHIDE PROSPECT IN KAMBALDA AREA, WESTERN AUSTRALIA: CASE STUDY

The limited geological outcrop within the Kambalda District, when coupled with the favourable physical properties of nickel sulphides, makes geophysical methods an important tool in the exploration for Archaean nickel deposits in this area. The present exploration strategy uses detailed airborne and surface magnetics in the targeting of favourable ore environments, structures, and prospective ultramafic–mafic contacts. Surface and downhole electrical and electromagnetic techniques are then applied to optimize prospect drilling and directly detect nickel sulphides. Thick, conductive overburden; magnetic ‘noise’ originating in near-surface laterites; ‘false’ anomalies due to conductive sedimentary units; and the extensive blanket of lake sediments in certain areas continue to present challenges to successful exploration. When exploration is focused upon areas close to known mineralization, surface geophysical methods may not have the required resolution to meaningfully aid geological methods. In such circumstances, downhole and underground surveys are used. Problems which can be addressed using such methods include:

- Isolating the geophysical response of closely spaced ore lenses.
- Discriminating the geophysical response of ore from that of nearby country rock.
- Evaluating the subsurface and hole-to-hole continuity of ore shoots.
- Elucidating subsurface structure by determination of fault displacements.

- Providing physical property information (typically magnetic susceptibility, natural gamma, apparent conductivity, and density) in order to correctly interpret surface geophysical responses. Applications in Kambalda-style nickel exploration include underground EM, underground seismic and radar, radio imaging, and downhole magnetometry.

Downhole electromagnetic (DHEM) is used to optimize prospect drilling for nickel in the Kambalda area (Trench and Williams, 2019). Survey geometry typically involves multiple transmitter loops at surface with $\frac{dB}{dT}$ responses read for successive receiver positions in hole. Occasionally, fixed-receiver EM (FREM; Coggon and Clarke, 1987) geometry has been used. To date, most DHEM applications have used a single-axis receiver probe, but three component works have recently been initiated. An example DHEM response reveals multiple anomalies due to conductive sedimentary units, known off-hole nickel sulphide mineralization, and a prospective off-hole anomaly. In this case, the anomaly of sediments decays more rapidly than those responses obtained from nickel sulphide. DHEM and physical property logging of drill holes confirm that the geophysical targets have been adequately intersected and/or indicate additional prospective locations.

REFERENCES

- Ali, T.H., Sas, M., Hood, J.H., Lemke, S.R. and Srinivasan, A., 2008. High speed telemetry drill pipe network optimizes drilling dynamics and wellbore placement. Society of Petroleum Engineers, 25 September 2012.
- The Australian Drilling, 1997. *Drilling: The Manual of Methods, Applications, and Management*. Boka Raton, FL: CRC Press, ISBN 9781566702423.
- Barron, J., Larson, B. and Shipboard Scientific Party, 1989. Ocean drilling program (ODP) leg 119. In: Barron, J. Larsen, B. and the Shipboard Scientific Party (eds), *Proceedings of the Ocean Drilling Program, Initial Reports*, vol. 119. College Station, TX: Texas A & M University.
- Bourgoyne, A., Millheim, K., Chenevert, M. and Young (Jr.), F.S., 1986. *Applied Drilling Engineering*. Richardson, TX: Society of Petroleum Engineers, p. 274, ISBN 1555630014.
- Coggon, J.H. and Clarke, E.H., 1987. The Fixed Receiver Electromagnetic (FREM) method for drillhole surveys. *Exploration Geophysics*, vol. 18, pp. 305–311.
- Darling, T., 2005. *Well Logging and Formation Evaluation*. Oxford, UK: Elsevier, p. 5, ISBN 0750678836.
- Dewan, T.J., 1986. “Open-Hole Nuclear Logging—State of the Art”—SPWLA Twenty-Seventh Annual Logging Symposium, June 9–13, Houston, TX.
- Enright, R.J., 1955. Sleuth for down-hole leaks. *Oil and Gas Journal*, pp. 78–79.
- Etnyre, L.M., 1989. *Finding Oil and Gas from Well Logs*. New York: Springer and Kluwer Academic Publishers, p. 249, ISBN 9780442223090.
- Fakhry, A.A., 2008. *Field Methods for Petroleum Geologists: A Guide to Computerized Lithostratigraphic Correlation Charts Case Study: Northern Africa*. Berlin and Heidelberg: Springer-Verlag and Springer Science & Business Media, p. 45, ISBN 9783540788379.
- Gluyas, J. and Swarbrick, R., 2004. *Petroleum Geoscience*. Oxford: Blackwell Publishing.
- Harald, B., 2012. Wireline Depth Determination, Rev 3.3, available via Society of Professional Well Log Analysts web site, www.spwla.org.
- Hilchie, D.W.D., 1990. *Wireline: A History of the Well Logging and Perforating Business in the Oil Fields*. Boulder, CO: Privately Published, p. 200.
- Kleinberg, L.R. and Jackson, A.J., 2001. An introduction to the history of NMR well logging. *Concepts in Magnetic Resonance*, vol. 13, no. 6, pp. 340–342, DOI: 10.1002/cmr.1018.
- McKinley, R.M., 1994. *Temperature, Radioactive Tracer, and Noise Logging for Well Integrity*. Washington, DC: U.S. Environmental Protection Agency (EPA), pp. 112–156.
- Prensky, S., 1999. Advances in Borehole imaging technology and applications. In: Lovell, M., Williamson, G., Harvey, P. (eds), *Borehole Imaging: Applications and Case Histories*, vol. 159. London: Geological Soc. Special Publications, pp. 1–43.
- Robinson, W.S., 1974. Field Results from the Noise-Logging Technique, 49th Annual Fall Meeting of the SPE of AIME in Houston. Texas. *Journal of Petroleum Technology*, vol. 28, no. 11, pp. 1370–1376. Paper Number: SPE-5088-PA.

- Sengel, E.W. “Bill”, 1981. *Handbook on Well Logging*. Oklahoma City, OK: Institute for Energy Development. p. 168, ISBN 0894191128.
- Sheriff, R.E. and Geldart, L.P., 1995. *Exploration Seismology*. 2nd edn. Cambridge: Cambridge University Press.
- Society of Professional Well Log Analysts, 1975. *Glossary of Terms & Expressions Used in Well Logging*. Houston, TX: SPWLA, p. 74.
- Stéphane, S., 2010. *Les diagraphies de corrosion*. Cachan Cedex: Ed. Lavoisier, p. 548.
- Taherdangkoo, R. and Abdideh, M., 2016. Application of wavelet transform to detect fractured zones using conventional well logs data (Case study: Southwest of Iran). *International Journal of Petroleum Engineering*, vol. 2, no. 2, pp. 125–139.
- Trench, A. and Williams, P.K., 2019. Nickel: Application of geophysics to nickel sulphide exploration in the Kambalda district, Western Australia. *ASEG Extended Abstracts*, 1994, vol. 1, pp. 169–180, DOI: 10.1071/ASEGSpec07_10.
- Wyllie, M.R.J., Gregory, A.R. and Gardner, G.H.F., 1958. An experimental investigation of factors affecting elastic wave velocities in porous media. *Geophysics*, vol. 23, pp. 459–493.

13 Airborne Geophysics

13.1 INTRODUCTION

Airborne geophysics is a powerful tool for studying huge areas and regions of the Earth and is found potentially very versatile for the regions located on hostile topography and that remained inaccessible to surveyors because of several unavoidable constraints related to coverage of the area or region under investigation. Since the early days of balloon photography and military reconnaissance, people have been struck by the broad view of the Earth that the airborne perspective provides. From the first photographs of the Earth taken from aircraft with handheld cameras to the highly refined swath terrain profiling systems now being developed, aircraft have provided a unique approach to studying Earth science. Compared with ground- or space-based methods, airborne techniques offer the advantages of improved access, rapid sampling at scales that are optimal for many geophysical problems, and a tremendous potential for interdisciplinary studies at intermediate scales. The following snaps may depict the setup of the airborne geophysical surveys conducted by mounting electromagnetic, gravity, and radiometric sensors on the aircraft (Figure 13.1).

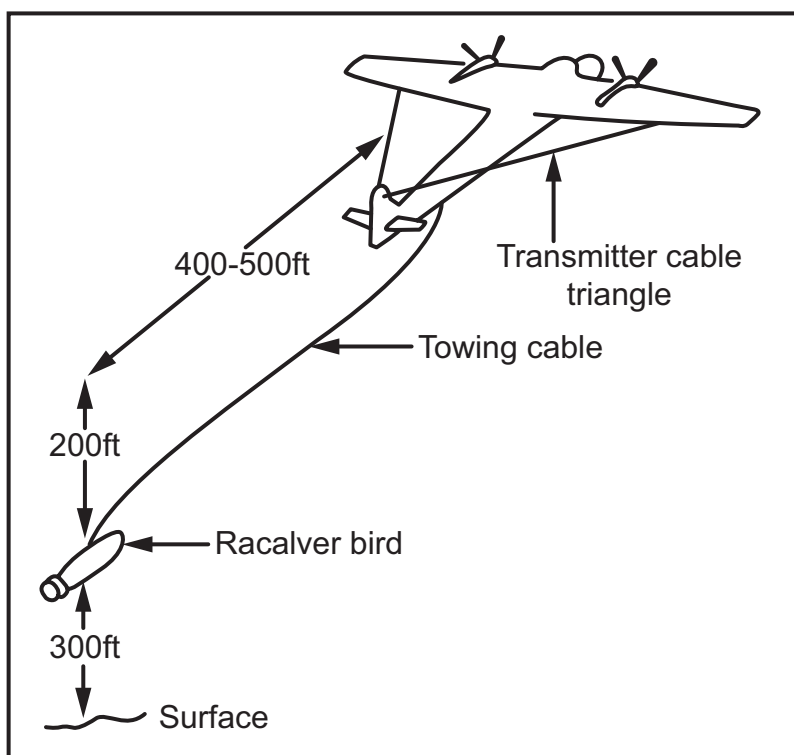


FIGURE 13.1 An illustration of an airborne EM system being used airborne.

13.2 ACCESS

Aircraft provide the capability of traversing regions that are otherwise difficult or impossible to undertake geophysical surveys using multisensors. Examples include remote areas of the arduous mountains, waterlogged places, and thickly vegetated area. Areas that are physically accessible but have social, economic, or political barriers are also potential candidates for an airborne survey. Where land access is restricted in environmentally hazardous areas because of health risks or industrial or military interests, the geochemical signatures of these regions can be mapped remotely. The land-based gravity survey is restricted to passable roads and stream valleys, whereas an airborne survey could systematically sample the entire region.

13.3 SAMPLING

Proper sample spacing is crucial for obtaining meaningful scientific results. Undersampling of a region may not allow resolution of the major features that control the system. Oversampling is inefficient and may restrict the extent of the area to be studied but may be necessary to isolate the signal from the noise. The sampling strategy chosen depends on the objectives of the survey. For example, magnetic anomalies that constrain rates of seafloor spreading can be adequately resolved with 20 km line spacing, aligned orthogonal to the ridge crest. In regions where plate motion is complex, a much closer line spacing (on the order of 4 km) with a complementary set of crossing lines is required to identify small-scale features and rotated crustal blocks. With airborne techniques, the scientific objectives of a study, rather than the restrictions of access, allow the optimal sampling strategy to be selected.

The improved sampling that is possible with an airborne program has particular advantages in the study of the potential fields of magnetics and gravity. Many of the errors and uncertainties that arise in the interpretation of land-based studies result from removing the effects of noise. For example, humanly generated magnetic fields introduce 'social' noise in magnetic studies, and surrounding, but poorly known, topography can introduce noise in gravity studies. Both of these measurement types are more easily interpreted if they are taken a systematic distance from the expected sources. To achieve optimal results, airborne gravity is typically collected at a single altitude, whereas airborne magnetic surveys are generally flown at a constant elevation above the terrain. A similar land-based survey would have required many more months of effort, as well as a large team of surface parties with diverse instruments and equipment to conduct the survey in areas with a variety of vegetation cover.

13.4 INTERDISCIPLINARY STUDIES

An important advantage of airborne survey systems is their potential for a flexible interdisciplinary platform. In the 1960s and 1970s, research groups collaborated on a systematic survey of the ocean floor using deep-water research vessels. The strategy was not simply to collect topographic measurements, to sample the ocean floor with cores, or to probe the structure of the ocean water column. Rather, these vessels provided a centre for integrated Earth science research. This approach has been the standard for oceanographic vessels since the *Challenger* expedition in the mid-1870s, which circumnavigated the globe and gathered a broad range of samples and observations. Alone, any one of the sample collections from this expedition would have been interesting, but together they provided a powerful stimulus for rethinking the structure and habitats of the globe. Aircraft can provide a similar centrepiece for interdisciplinary research. The capabilities aircraft offer for such research can be exemplified as follows. In many regions, the seepage of saline groundwater into agricultural areas adversely affects crop production. In order to determine groundwater flow patterns, aircraft could carry instrumentation for both airborne magnetic surveys and electromagnetic mapping. The airborne magnetic data would delineate the geologic structure, and the electromagnetic data would delineate the extent of saline groundwater. Combining these two sets of data may shed light on the flow patterns and their control in the presence of a series of igneous dykes that

cut across the area. This capability of airborne multiple instrumentations may provide new insight into the linked processes that control the Earth's systems. The major drawback of using multi-instrumented aircraft is that such an endeavour is economically not so much viable, which can be overcome by adopting a measure of compromising individual measurement types with as minimum as possible cost of airborne geophysical surveys. As noted earlier, the optimal altitude for collecting airborne gravity and magnetic measurements may differ. Nevertheless, in many instances, such interdisciplinary efforts, coupled with new high-resolution techniques, do offer increasing opportunities for airborne geophysics. Estimates of precise positioning of target locations based on GPS technology will be increasingly in use under the advanced applications of airborne surveys in search of new reserves along with their diverse applications.

13.5 PRECISE POSITION CAPABILITIES AND REQUIREMENTS

Airborne geophysical experiments require that the location of the measurements be accurately known and, in some cases, that the orientation of the platform from which the measurements are to be taken be well constrained. Accurately locating the experiment is the real-time, or navigational, problem. In large-scale regional studies, real-time navigation is of minimal concern, but as the applications of airborne geophysical techniques become increasingly detailed, and as the instrumentation becomes increasingly sensitive, navigation issues become more significant. Navigation issues are critical, for example, in monitoring changes in topography after an earthquake. The scientific objective of understanding the deformation of the Earth requires that the same track be flown both before and after the seismic event. Recent applications of pseudosatellites have produced encouraging results in very precise, real-time applications and could have application in postprocessing. Most geophysical applications require postprocessing of the data because of our inability to hold the aircraft to within a metre. In general, postmission positioning requirements are more stringent than are real-time navigation requirements, and thus, more sophisticated algorithms are used for aero triangulation applications, for example, or in computing aircraft acceleration for gravity applications. Recovering the aircraft orientation is important to many applications, as this information may be required to correct for aircraft motion during data acquisition. Examples are the correction of laser ranges to the vertical and also the rectification of imagery. The accuracy required for navigation, positioning, and attitude varies among operations, but in general, the exploitation of data obtained from airborne sensors increases with improved positioning and navigational capabilities.

GPS is a powerful tool for determining aircraft positioning and navigational attributes and is potentially useful for recovering orientation. The system has the advantages of being globally accessible, little affected by local weather conditions, and having relatively low user costs. The GPS system is based on a constellation of 24 satellites that transmit biphasic encoded signals at two frequencies, denoted by *L1* (1.575 GHz [gigahertz]) and *L2* (1.227 GHz). Two basic codes are written on the GPS carrier signals at two frequencies:

1. The coarse/acquisition code (C/A code)
2. The precise code (P-code)

The system was designed to provide navigation to an accuracy of approximately 10 m with the P-code, and approximately 100 m with the C/A code. Researchers accessing the full range of signals transmitted by GPS satellites have expanded the capabilities for positioning moving vehicles (such as aircraft) to the decimetre level using differential techniques. The three main types of observables that are used in the analysis of GPS data to position objects precisely are:

1. The pseudorange
2. The carrier phase
3. The Doppler shift

Pseudorange measurements are made by determining the difference between the arrival time of a GPS signal (as measured on the receiver clock) and its transmission time (as determined by the satellite clock). By simultaneously observing four or more satellites, it is possible to determine the position of the receiver and to correct for differences in time between the receiver's clock and the GPS satellite time system. When selective availability (SA) is turned on, there is high-frequency 'dithering' of the satellite clock by up to 0.2 μ s (microseconds). This dithering cannot be recovered from the ephemeris message without access to a classified decryption key. This noise introduces 60 m errors into the pseudorange solutions with periods of several minutes.

Carrier phase measurements are made by reconstructing the carrier signal, the fundamental frequencies of *L1* (1.575 GHz) and *L2* (1.227 GHz). This process includes removing the biphasic encoding and measuring the phase difference between the reconstructed carrier phase and a local oscillator within the receiver for both the *L1* and *L2* frequencies. As this carrier-beat phase rotates through cycles, the number of cycles is accumulated. Thus, the phase measurement is the accumulated phase change from the time the satellite is acquired by a receiver (locked on) until the receiver loses the signal (loss of lock). The carrier phase noise is a few millimetres; however, it also contains an ambiguity that must be determined to exploit this accuracy.

The *Doppler observable* is most often determined from the time derivative of the carrier phase. The Doppler observable is rarely used for geodetic positioning, but it can be useful for connecting phase measurements across small gaps (<5 s), when the signal from a satellite is lost (e.g. during banked turns or rapid accelerations). Together, these three observables provide a powerful suite of tools for determining the time and position of a vehicle anywhere around the globe.

The principal errors in a GPS position arise from errors in timing and errors introduced as a result of the propagation of the signal from the satellite to the receiver. The SA signal is a clear example of an artificially induced error that affects the pseudorange measurements and thus position determinations. The SA error size is much larger than the timing errors that are intrinsic to the GPS clocks. The propagation errors are associated with dispersive delay in the ionosphere, atmospheric delays, and the signal reflecting off objects surrounding the antennas before it is recorded by the receiver (multipath).

Differential techniques based on multiple satellites and receivers are used to minimize many of these errors. Differential techniques are based on the concept that many of the errors in timing and propagation can be eliminated by using a reference receiver in a fixed location and positioning the moving receiver relative to this reference. This configuration assumes that systematic errors seen by both receivers can be removed by differencing the signal. The accuracy of differential positioning decreases with increasing baseline lengths. This approach can be applied to both pseudorange and carrier phase measurements. Differential pseudorange positions are accurate to 0.5 to 3.0 m, depending on the receiver type used, and are being used experimentally for commercial aircraft systems. Differential carrier phase measurements are as accurate as 2 to 20 cm and generally require extensive postmission analysis, although real-time systems are under development.

Several efforts have been made to demonstrate the robustness of the differential GPS approach for accurately locating aircraft. As the most difficult component to constrain is the vertical position, the recovery of this component is generally used as a benchmark. A ranging system that provides an independent constraint on the aircraft height is used in combination with precise GPS positions. In experiments over water or ice, the position of the aircraft recovered from the independent, and probably more accurate, ranging system is within 20 to 50 cm of the elevation measurements recovered from the differential GPS carrier phase measurements. Other airborne tests have been conducted using inverse photogrammetry, in which the camera positions are independently determined from aero triangulation with ground control and then are compared to the estimated GPS positions.

13.6 MEASURING THE EARTH'S GRAVITY FIELD

A striking example of the benefits of precise positioning of a research aircraft is the determination of the Earth's gravity field. Accurate knowledge of the Earth's gravity field is of major importance in geodesy and geophysics. In *geodesy*, the gravity field defines a reference, the geoid (or mean sea level under ideal conditions), for topographic heights (Heiskanen and Moritz, 1967). The geophysical utility of surface gravity data lies in constraining the density structure of the Earth. Furthermore, the Earth's gravity field is one of the few quantities (magnetics and topography being the others) that contain the same levels of spatial detail as the Earth's density and can be measured and mapped at all these levels by a variety of techniques. Gravity measurements have applications over a broad spectrum of scales, ranging from tens of thousands of kilometres to less than 1 km. The accuracy needed, corresponding to the strength of the signal, ranges from parts per hundred thousand to better than a part per million. With just a few global Earth parameters, the gravity field can be modelled to an accuracy of a few parts per 10,000; the remaining structure on global, regional, and local scales must be measured. On land, measurements of relative gravity historically have come from static deployments of the pendulum, the torsion balance, and (since 1950s, exclusively) the spring gravimeter (Torge, 1989). Coverage with these types of instruments is limited inherently by geographic accessibility, a stable environment, and human and economic resources. However, given the resources and sufficient time, land gravimetric and, to some extent, sea gravimetric could be measured at high resolution over many parts of the world. In fact, large gravimetric databases exist for most of the world's land areas, although coverage is not uniform. Some parts of Africa, South America, Asia, and the polar regions are characterized by no surface data at all. The global structure of the gravity field can be discerned by tracking an Earth-orbiting satellite from several ground stations and observing the orbital perturbations caused by the Earth's asymmetric mass distribution. Today, with the accumulation of tracking data from numerous satellites, and with constantly improving tracking techniques, the Earth's gravity field can be derived from these data alone to wavelengths greater than about 500 to 600 km (Nerem et al., 1994). Over the oceans, satellite altimetry yields measurements of the geoid with horizontal resolutions of tens of kilometres (Rapp and Pavlis, 1990), although with insufficiently uniform accuracy at these short wavelengths for many geophysical applications. Increasingly accurate measurements from satellite altimetry (less than 10 cm for the current TOPEX/POSEIDON satellite) characterize the instantaneous sea surface height, which can deviate from the geoid by as much as 1 to 2 m. Therefore, satellite altimetry is not the final answer to better marine gravity models. Gravity measurements made from a moving vehicle, such as an aircraft or ship, are contaminated by motion-induced and inertial accelerations due to the rotating coordinate frame, even when mounted on a stabilized platform. These accelerations must be accurately determined and removed from the measurements to recover meaningful gravity anomalies. The amplitude of the vertical acceleration can be much greater than that of the geological and geophysical significant anomalies. For example, a sedimentary basin may have a gravity anomaly of 125 mGal, whereas the typical vertical accelerations of an aircraft are approximately 20,000 mGal. Also, the Eötvös correction (Coriolis and centrifugal effects), being a function of flight direction and velocity, can reach amplitudes of over 1,000 mGal and must be calculated for gravity measurements. On ships, the requirements on position and velocity accuracy are less demanding because of the lower speeds and the somewhat-predictable average vertical position. These less-stringent positioning requirements have permitted the widespread use of marine gravimetric through the use of satellite positioning and radio navigation systems, such as Loran-C. Early airborne gravimeter measurements also used Loran-C, as well as radar, laser, and barometric altimetry. These techniques, however, were largely imprecise and regionally limited. It was not until the late 1980s that precise kinematic positioning with GPS was shown to meet the rigorous positioning requirements for airborne gravity measurements. GPS positioning, which is

easily made on global scales, has led to the broader application of airborne gravity surveys, both in scientific research and in commercial exploration. Numerous basic and applied research groups in the United States, Canada, Switzerland, and Germany are actively pursuing both proof of concept and operational airborne gravity capabilities that are now possible with GPS.

13.7 MEASURING THE EARTH'S SURFACE TOPOGRAPHY

Measurements of surface elevations are fundamental to many applications of geology, geophysics, hydrology, terrestrial ecology, geomorphology, glaciology, and atmospheric physics. For example, the topography of the polar ice caps and mountain glaciers is important because it is a direct measure of ice dynamics and is closely linked to global climate and sea level change. Existing topographic data in 'well-mapped' North America, Western Europe, Australia, and in mineral blocks located in Rajasthan and other regions in India, however, are inadequate in terms of accuracy and consistency to support most types of scientific research. For example, in high-relief terrain, the vertical accuracy of available data may be worse than 30 m, too poor by at least an order of magnitude to support many applications. Less-accessible regions of the Earth, including large areas of Africa, Asia, South America, and Antarctica, do not have even this level of topographic coverage. Topography of the survey sites has traditionally been measured by triangulation and levelling methods. This land-based effort required skilled crews to measure accurately distances and elevation changes. Over the past several decades, ground determinations of topography have been supplemented by airborne photogrammetry, which provides the potential to map the Earth's surface to a decimetre. Photogrammetric techniques are adequate to meet topographic mapping standards for scales of 1:10,000 or less, but they are not sufficiently precise to monitor dynamic changes in the Earth's surface. The signals of interest to Earth scientists are typically on the order of millimetres to tens of centimetres for postseismic rebound, for example, millimetres for stream erosion rates, and tens of metres for mass wasting processes. Precise positioning is required for these applications.

With the implementation of geodetic networks supported by land-based GPS crews, points can now be located on the order of millimetres. Recent studies have documented plate motions and have raised important questions about the nonrigid nature of the Earth's deformation along plate boundaries. A drawback of this technique, however, is that the geodetic networks provide only point measurements, and knowledge of the entire 3-D surface is required if Earth processes are to be understood. The recent integration of airborne mapping techniques with precise positioning has demonstrated the ability of this approach to detect dynamic processes.

13.8 VERY HIGH-RESOLUTION STUDIES WITH AIRBORNE TECHNIQUES

Land-based gravity measurements have long been used to identify the salt domes and other structural traps for oil along the Gulf Coast and the large mineral deposits in the Western United States. As the easily identified reservoirs and mineral deposits become depleted, however, exploration industries will require increasingly detailed information on the subsurface of the Earth. High-resolution mapping of the subsurface is also rapidly developing as a requirement for the characterization and clean-up of superfund and other waste sites. Measurements of these waste sites often pose health hazards to ground crews, so there is great interest in the development of remote tools. These high-resolution applications can be addressed by airborne geophysical technology and precise positioning. An innovative approach was developed utilizing a helicopter equipped with GPS and airborne electromagnetic, magnetic, and radiation sensors.

The Geological Survey of India (GSI) has been carrying out multisensor aerogeophysical surveys in different geological terrains of India since 1965. To date, about 23.88 lakh km² area, with around 15 lakh line km, has been covered with various survey specifications under different projects. The study of aerogeophysical surveys has led to the discovery of mineral deposits, like the Kayar and Aldahalli in Karnataka, and Golapalle in Andhra Pradesh. The survey has also delineated significant

uranium in Nalgonda, Andhra Pradesh. In the first phase, magnetic-gradiometric and radiometric surveys are taken up by engaging project implementing agencies (PIAs) for data acquisition, processing, integration, and interpretation. It will be followed by high-resolution EM, gravity, gravity gradiometric, and magnetic surveys in selected smaller blocks. The government wants to identify mineral potential areas and attract private investment in mineral exploration using geophysical data.

13.8.1 AIRBORNE ELECTROMAGNETIC METHOD

In the 1950s, ground and airborne electromagnetic (EM) surveys became the primary field method preferred by geophysicists in the exploration for massive sulphides in most of Canada. However, there existed an overarching shortcoming. The limitation that effected electromagnetic methods was depth of exploration. For airborne surveys, it was limited to only 75–200 ft, equivalent to approximately 20–60 m (Ward, 1959). It was both the signal-to-noise ratio and signal strength that limited the usefulness of this exploration method.

In 1955, L.B. Slichter proposed that natural electromagnetic sources, like lightning strikes associated with thunderstorms, could be utilized as a source of energy to replace artificially transmitted energy used in conventional EM methods of the time. The hope was to avoid the issues associated with poor depths of exploration (Slichter, 1955). It was Ward, in 1959, who concluded that it may be practical to utilize audio frequency magnetics (AFMAG) as an exploration tool and first demonstrated its ability to locate and identify buried conductive bodies (Ward, 1959). Ward essentially established a relationship that characterized AFMAG as a scalar quantity whereby

$$H_z = T_x H_x \quad (13.1)$$

Eq. 13.1 states that the vertical magnetic field H_z is related to the horizontal magnetic field H_x through a transfer function T_x , the tipper. T_x is the component of the tipper vector in the horizontal direction. Despite many positive attributes, poor-quality data and an inability to repeat successful trials became major stalling points to both AFMAG's application and acceptance in the mineral exploration industry (Ward et al., 1966, 1968).

By the end of the 1960s, Ward and his colleagues had spent ten years investigating and experimenting with the AFMAG method. They determined AFMAG to be a special-purpose tool that could be effectively used in reconnaissance surveys (Ward et al., 1966). They noted that the application seemed promising as a geological mapping tool. It had been hoped that the product could be used to directly identify massive sulphide ore bodies, but certain peculiarities of the method had reduced its application in this regard (Ward et al., 1966). These peculiarities included:

1. The natural alternating magnetic fields in the 150 Hz and 510 Hz frequency bands employed in typical AFMAG systems show significant seasonal and diurnal variations in field strength which prohibit year-round application and even curtail midday application at times of the year when the fields are strongest. This effect is strongly latitude-dependent, the strongest fields existing in a relatively narrow belt about the equator.
2. Local atmospheric discharges, especially lightning, produce a natural alternating magnetic field of grossly erratic amplitude and direction which renders measurements difficult or impossible.
3. Power line fields are frequently much greater than natural fields and hence limit the area of applicability.
4. The dip angle and azimuth measurements are not strictly reproducible so that results obtained one day do not agree with those obtained another day.
5. Topographic relief alone leads to anomalies.
6. It is difficult or impossible to map conductors whose axes are nearly parallel to the direction of the inducing field.

7. False anomalies arise, especially with the airborne version of AFMAG.
8. Faults, shears, and carbonaceous horizons are delineated by AFMAG so that it is very difficult to locate sulphide bodies because of their small relative response.
9. The dual-frequency interpretation is not capable of distinguishing conductors of different (conductivity $\times$ thickness) products.
10. Some sulphide ore bodies do not give a measurable response on the AFMAG system, especially the airborne version.
11. AFMAG ground surveys are very laborious.
12. The system responds to broad surface resistivity variations.

However, it was determined that AFMAG was a preferential method for deep exploration, mapping faults, locating sulphides buried beneath deep glacial ice, and exploring in rugged terrain (Ward et al., 1966). More field, experimental, and theoretical studies were clearly still required to fully describe AFMAG. In fact, the scalar estimation proposed by Ward (eq. 13.1) was proven to be incorrect in the early 1970s. In 1972, Keava Vozoff correctly redefined the relationship between the vertical and horizontal components of the magnetic field and the tipper tensor to be

$$H_z(f) = \{T_x(f) \times H_x(f)\} + \{T_y(f) \times H_y(f)\} \quad (13.2)$$

where f is the frequency; H_x , H_y , and H_z are the components of the magnetic field in three orthogonal directions; and T_x and T_y are the two horizontal components of the tipper tensor. Eq. 13.2 is the basis of the AFMAG exploration method. Vozoff made it clear through his research and experimentation that a large amount of valuable geological information could be obtained if the AFMAG method was applied carefully (Vozoff, 1972). At this time, the tools were available to perform reliable surveys. However, survey design was often based on a preliminary investigation of the site. A correct precursory evaluation was required in order to return a technically sound and economically viable survey (Vozoff, 1972).

Research and development of AFMAG slowed in the 1970s through the 1990s. One exception was the work of Labson et al. (1985). Labson and his colleagues further refined the work of Vozoff. They made improvements to the AFMAG method and developed a spectral, Fourier-domain-derived tipper measurement that was repeatable on a year-round basis and was independent of the polarization state of the natural source field (Labson et al., 1985). Their introduction of a remote referencing technique for tipper measurements also made high-quality data acquisition possible using portable sensors. This allowed the technique to be applied to time periods of low signal strength. The group also began utilizing a broad band of frequencies which made this method feasible and effective for use in areas where depths of exploration may range from several metres to several kilometres (Labson et al., 1985).

However, once again, the airborne AFMAG method continued to be set aside in favour of inductive EM systems. Geotech Ltd. re-established AFMAG development in 2000, using the signal processing proposed by Labson et al. (1985). Their efforts eventually led to a technology named ZTEM (Z-Axis Tipper Electromagnetics) in 2005. The system was brought into operation in 2006 and represents the first commercial airborne AFMAG system available in 30 years (Legault, 2012).

13.8.1.1 Development and Description of the ZTEM System

In 2000, Geotech Ltd. re-embarked on the development of the AFMAG technology with the knowledge of these limitations. Between 2000 and 2005, several systems were established (Figure 1, p. 6 in Connors, 2013: ZTEM system in 2006, airborne receiver in foreground with 7.4 m base stations coils deployed in background; modified after Kuzmin et al., 2005). It was hoped that the application of modern technology and new digital signal processing techniques would allow the development of a new commercially viable EM system that reduced or eliminated some of the deficiencies that Ward and other researchers had previously identified (Kuzmin et al., 2005).

Test results were positive, but shortcomings were identified. These included receiver coils that did not possess high enough sensitivity and a suspension system that did not effectively dampen vibrational noise. Funding from industry and government allowed Geotech Ltd. to significantly upgrade the system hardware. This included new, much more sensitive receiver coils, a new suspension system, upgraded base stations, and orientation sensors. Numerical simulations, field trials, data processing, and reporting techniques were also established. The technology was patented in 2002 (Legault, 2012).

Following the success of these improvements, further additions to the AFMAG system followed. The electronics and suspension systems were further modified, and a tail and drag skirt were added to increase the airborne stability of the platform. Ongoing ground and air testing identified a need to further develop the suspension system and receiver size in order to reduce the signal-to-noise ratio (Figure 2, p. 7 in Connors, 2013: Current ZTEM suspension system and airborne receiver, possessing a diameter of 7.4 m; modified after Kuzmin et al., 2005). The distance between the aircraft and the receiver was also maximized by implementing the use of a longer tow cable (90 m).

The use of base station receivers to capture horizontal field data resulted in accurate and interpretable data and permits the use of a single coil to measure the Z field (H_z) from the aircraft. Other benefits to this approach included the ability to increase the size and sensitivity of the base stations and airborne receiver coils (Kuzmin et al., 2005). This allowed the base stations to be placed well away from man-made structures, further improving their signal-to-noise ratio. These experimental innovations eventually led to the successful development of the Z-Axis Tipper Electromagnetic system (ZTEM) (Figure 3, p. 8 in Connors, 2013: ZTEM airborne receiver in foreground with 3.5 m base stations coils deployed in background; modified after Kuzmin et al., 2005).

ZTEM is an airborne geophysical system that was first flown by Geotech Ltd. in 2006. It is a passive electromagnetic system, meaning, it does not require a man-made transmitter. The ZTEM system detects electromagnetic responses induced by worldwide thunderstorm activity. These storms are often located thousands of kilometres (km) from the survey area. ZTEM data can be used to map the resistivity/conductivity of the subsurface. Depending on the signal strength, the ZTEM system can detect a frequency in the 'audio range' from 30 Hz to 720 Hz. Lower frequencies are especially effective for penetration through conductive overburden. The ZTEM system is an excellent deep exploration tool sensing depths down to 2,000 m and provides excellent resistivity mapping and identification of weak anomalies.

With the ZTEM system, the magnetic tensor $T = (T_{zx}, T_{zy})$ is known as the tipper. The components of the tipper are frequency-dependent complex numbers. The primary field is assumed to be uniform and horizontal. It is measured by the base station receivers (Figure 4, p. 9 in Connors, 2013: ZTEM base-station receiver coil in 2007 (3.5 m); modified after Kuzmin et al., 2005). Lateral contrasts in the conductivity of the basement material cause the appearance of a vertical component. This vertical component is sensed by the airborne receiver.

The original version of ZTEM utilized three on-board orthogonal receiver coils contained in a towed 'bird'. This configuration has since evolved with the ZTEM system. Presently, the system has been modified to use a larger single vertical air coil and fixed base station coils. Base station receiver coils are deployed perpendicular to one another. Originally, the two base station coils had square dimensions with 7.4 m lengths. However, these larger coils were replaced by similar coils with square dimensions of just 3.5 m. This change allowed more versatility and ease of deployment in remote or difficult terrain. In 2009, the base station receivers designed for Geotech's air MT technology were adapted for ZTEM (Figure 5, p. 10 in Connors, 2013: ZTEM Air MT base-station receiver (3.0 m), often adapted for use with ZTEM; modified after Legault et al., 2012). These base station coils have even smaller square dimensions of only 3 m and are even easier to deploy, which can often be done with a helicopter.

ZTEM utilizes a passive airborne receiver. The coil dimensions are 7.4 m in diameter (Figure 6, p. 11 in Connors, 2013: ZTEM system in flight, see labels for dimensions and configuration details; modified after Legault et al., 2012). It is oriented horizontally, producing a vertical magnetic dipole.

The airborne receiver is isolated from flight and aerial vibrations by a patented suspension system. The entire apparatus is enclosed by a fibreglass shell and is towed by a helicopter using a 90 m long cable. The optimal flight altitude for the sensor is 80 m above the ground surface (Figure 6, p. 11 in Connors, 2013: ZTEM system in flight, see labels for dimensions and configuration details; modified after Legault et al., 2012). The cable used to tow the receiver also doubles as an antenna used to transmit the measured EM signals back to the on-board data acquisition system.

13.8.1.2 Estimation of Depth of Exploration (Skin Depth)

With respect to numerical values, it is expected that the ZTEM sensor response will decrease as the frequency under investigation increases. This hypothesis is based on the theory used to investigate the depth of exploration for current magnetic systems. A combination of the laws of Ampere and Faraday have been used to derive a diffusion equation that describes the propagation of electromagnetic fields in the Earth. When EM fields propagate into the subsurface, their amplitude is attenuated, and they are delayed in time. The depth of exploration (skin depth) can be calculated for magnetotelluric exploration:

$$\delta \text{ (km)} = 0.503 \sqrt{\rho T} \quad (13.3)$$

where the exploration depth (skin depth) is δ (km), ρ (Ω -m) is resistivity, and T is period (s). Eq. 13.3, if applicable to ZTEM, suggests that EM techniques are insensitive to conductivity anomalies that lie greater than 1 skin depth deep. This also suggests that lower frequencies will result in greater resolution at depth (Figure 7, p. 14 in Connors, 2013: Illustration demonstrating attenuation of amplitude with increasing frequency; adapted from M. Unsworth, U. of Alberta; modified after Legault et al., 2012).

13.8.1.3 Use of Type Curves to Model AFMAG and ZTEM

An interpretation regime for AFMAG was first proposed by Ward et al. in 1968. They analyzed magnetic fields scattered from thin disks and constructed interpretation diagrams for the circular disk scenarios. From these diagrams, one was able to estimate field data that included dip, depth of burial, radius, and the (conductivity $\times$ thickness) product (Ward et al., 1968). Characteristic curves, also known as type curves, are common in academic literature and have been used for interpreting dipping sheet conductors (Telford et al., 1976). This has not been the case for AFMAG. However, during the 1970s, the introduction of numerical finite element, finite difference, and integral equation 2-D forward model solutions for plane wave methods (Hohmann, 1987) has allowed for the calculation of tipper responses for AFMAG interpretation. This is most likely why there is a lack of graphic-based interpretation tools. It was the work of Sattel and Witherly, in 2012, that first presented examples of 2-D- and 3-D-type curves for ZTEM data.

13.8.1.4 Modelling Software: Occam 2-D MT

The OCCAM 2-D MT software is available to compute synthetic ZTEM sensor responses. The software also supports many data types, including apparent resistivity, phase, impedances, and magnetic field tipper values. The software package directly mimics the response that would be measured by the actual ZTEM sensor (i.e. mobile H_z field with fixed H_x field measurements).

In terms of airborne EM equipment development, the AEM method has been used in China since the 1970s. The time domain AEM system (M-1), which was designed by the Changchun College of Geology, has been used already in 1970 (Yin et al., 2015). During 1981–1983, the time domain AEM system (M-2) was developed from the M-1 system; unfortunately, the research was stopped because of a lack of funding. Since 2000, the fixed-wing triband aviation electromagnetic system HDY-402 was developed by the Geophysical and Geochemical Exploration Institute of the Chinese Academy of Geological Sciences, while the CHTM time domain helicopter aerial electromagnetic

survey was developed by the Land Resources Aviation Geophysical and Remote Sensing Center (Yin et al., 2015). The HDY-402 system uses an upright coplanar transmitter and receiver with a transmission and reception distance of 19.2 m; operating frequencies of 463 Hz, 1,563 Hz, and 8,333 Hz; a maximum sampling rate of eight times per second; and a noise level of less than 20 ppm. When the zero drift is less than 100 ppm/hour, the exploration depth is about 100 m. With increasing altitude, the intensity of the signal gets weaker; hence, the application of this system is limited to detecting shallow ore bodies and groundwater resources (Yin et al., 2015).

13.8.1.4.1 Case Study: Widgie Townsite Nickel Prospect, Western Australia

Conductive regolith and saline lake sediments within the Kambalda district limit the widespread application of time and/or frequency domain airborne electromagnetic (AEM) systems for the direct detection of nickel sulphides at depth. However, AEM has proved efficient in mapping variations in regolith conductivity thickness and in delineating groundwater resources in the Kambalda area, Western Australia. There exists a relationship between regolith geometry and bedrock geology (Trench and Williams, 2019); the method is also capable of identifying geophysical targets requiring follow-up surface work. The figure in Trench and Williams (2019) illustrates examples of regolith and shallow bedrock AEM anomalies (GEOTEM system, 75 Hz base frequency, 4.9 ms latest channel) in the vicinity of the Widgie Townsite prospect. In general, however, nickel sulphide prospects in the Kambalda district are too deep to give an AEM response discernible from that of the conductive cover.

13.8.2 AIRBORNE MAGNETIC METHOD

The first geophysical airborne research method was the magnetic method. Discovered by Faraday, the method was initially employed by the USSR (currently Russia) in 1936 (Hood and Ward, 1969) and adapted later with better modifications by the United States in 1940 (Hood and Ward, 1969). Both countries had vested interests in military and technology, particularly for submarine applications. On some of the improvements, another early airborne survey was made in the United States in 1944 using the Beech Staggerwing NC18575 (Morrison, 2004).

An aeromagnetic survey is a common type of geophysical survey carried out using a magnetometer aboard or towed behind an aircraft. On-board systems are known as stinger systems and are typically installed on the tail of the aircraft (Camara and Guimaraes, 2016). The aircraft is then adapted to prevent any materials from influencing the measurements. For example, when conducting magnetic measurements, the aircraft is assembled with the least possible number of metallic substances or surfaces. Sensors are typically installed in the aircraft's extremes, such as the wingtips, so that mechanical or human factors do not affect the measurements. Pilots must be wary of the performance loss caused by the addition of sensors to the wingtips as they affect aerodynamics. The principle is similar to a magnetic survey carried out with a handheld magnetometer but allows much larger areas of the Earth's surface to be covered quickly for regional reconnaissance. The aircraft typically flies in a grid-like pattern, with height and line spacing determining the resolution of the data (and the cost of the survey per unit area).

As the aircraft flies, the magnetometer measures and records the total intensity of the magnetic field at the sensor, which is a combination of the desired magnetic field generated in the Earth as well as tiny variations due to the temporal effects of the constantly varying solar wind and the magnetic field of the survey aircraft. By subtracting the solar, regional, and aircraft effects, the resulting aeromagnetic map shows the spatial distribution and relative abundance of magnetic minerals (most commonly the iron oxide mineral magnetite) in the upper levels of the Earth's crust.

Airplanes are normally used for high-level reconnaissance surveys in gentle terrain, and helicopters are used in mountainous terrain or where more detail is required. Magnetic fields are measured using magnetometers. A single sensor is normally used to measure the magnitude of the Earth's magnetic field, or the magnetic field intensity (in units of nanotesla: nT), but two sensors mounted

either vertically above or horizontally adjacent to each other can be used to derive the vertical or horizontal magnetic gradient (in units of nanotesla per metre: nT/m). The advantage of gradiometers is that temporal variations in the magnetic field are essentially cancelled out by subtracting the signal from one magnetometer from the signal of the other (www.reynolds-international.co.uk, 'Technical summary sheet no. 1'). Systems with off-board equipment typically carry the magnetic sensor, often called the bird, below the plane. This requires precise flying and a high level of compensation to attain reliable data. An example of the system off-board is shown in the figure in Camara and Guimaraes (2016). Aeromagnetic surveys are also used to perform reconnaissance mapping of unexploded ordnance ('Aeromagnetic Survey', Wikipedia, 21 April 2020). The aircraft is typically a helicopter, as the sensors must be close to the ground (relative to mineral exploration) to be effective. Electromagnetic methods are also used for this purpose. The measured magnetic field is the resultant of the Earth's magnetic field and the anomaly field. The shape and amplitude of a magnetic anomaly depend on the location, shape, size, strike, burial depth, magnetic susceptibility, and intrinsic magnetization of the causative body, and the angle at which the survey lies relative to both the Earth's magnetic field and to the causative body. Additionally, the Earth's magnetic field varies in time on a scale of minutes to centuries. Typically, a normally magnetized object (one with no intrinsic field of its own) located in the Northern Hemisphere produces a characteristic N–S anomaly (Reynolds, 2011, 'Technical summary sheet no. 21').

During airborne geophysical acquisitions, it is necessary to conduct data quality checks. In general, quality assessment and quality control (QA-QC) are conducted on all potential methods of geophysics and gamma spectrometry, which are limited by lateral offset and acquisition speed. Parameters that undergo QA-QC analyses include the magnetic field, temperature, spectrum range, and others. In addition, the acquisition area and control area are generally broken into grid blocks. The figure in Camara and Guimaraes (2016) illustrates the quality of two types of data acquisition. The figure in Camara and Guimaraes (2016) was measured during the 1980s, when measurement equipment was much less sophisticated. The figure in Camara and Guimaraes (2016) showed a set-up with a 250 m line spacing and using the latest equipment. Both refer to the same area, located in the southern portion of the Minas Gerais State, Brazil. Although a complete database was unavailable for Camara and Guimaraes (2016), the observed level of detail is much higher than in the figure (Camara and Guimaraes, 2016).

Note that developments in the airborne geophysics field have led to exponentially improved data, in terms of both coverage and quality. These data have allowed for significant mineral exploration, geological studies, and geophysical analysis in Brazil and across the globe. For example, the figure in Camara and Guimaraes (2016; Guimaraes et al., 2014) illustrates a subsurface map of high-resolution aeromagnetic data, where the degree of certainty decreases as the data resolution increases. Geologic structures (like ore bodies or faults) often produce small magnetic fields that distort the main magnetic field of the Earth. These 'anomalies' can be detected by measuring the magnetic field near the surface of the ground. By analyzing these measurements, geophysicists can learn about geologic structures, even though the structures may be concealed entirely below the Earth's surface (<https://pubs.usgs.gov/of/1999/of99-514>, maintained by Michael F. Diggles, last modified 27 October 2005, USGS about aeromagnetic survey). Other studies show results of these evolution processes of the airborne data geophysical acquisition. A few examples in the scientific literature include Ravat (1996), LaBrecque and Ghidella (1997), Brozena et al. (2002, 2003), Finn and Morgan (2002), Salem and Ravat (2003), Hinze et al. (2005), Hemant et al. (2007), Bouligand et al. (2014), and Guimarães et al. (2014).

13.8.2.1 Aeromagnetic Data over the Komatiite-Hosted Nickel Sulphides in the Kambalda District, Western Australia: Case Study

The nickel sulphide deposits of the Kambalda district comprise part of a richly mineralized area of Archaean granite–greenstone terrain in the southern part of the Eastern Goldfields Province of the Yilgarn Craton, Western Australia. The Kambalda district refers to the NNW trending segment of

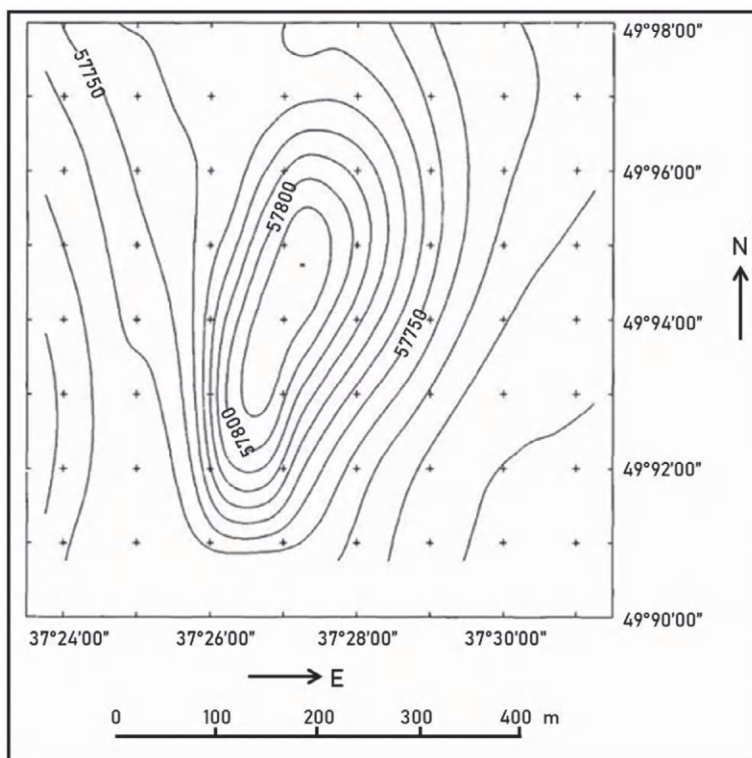


FIGURE 13.2 Contours of TMI over the Mariners nickel sulphide deposit reveal an elongate magnetic high over the ore zone. The deposit is ‘blind’ as it is covered by Tertiary lake sediments. (Aerodata multichannel survey, 200 m line spacing east–west, 60 m flight height.) Contour interval is 10 nT. The strike of the basal contact and mineralization is north–south and parallels the elongation of the magnetic anomaly.

Source: Trench and Williams (2019).

the Norseman–Wiluna belt between Kalgoorlie and Norseman (Trench and Williams, 2019). The geology of this area has recently been summarized by Griffin (1990) and Swager et al. (1990), and only those factors pertinent to geophysical exploration are noted here. Many of the exploration and mining leases in the Kambalda area have now been covered by several generations of aeromagnetic data. The survey specifications of these datasets vary from reconnaissance level (1.6 km line spacing, 150 m flight height) to ultradetailed (50 m line spacing, 50 m flight height). Recent fixed-wing surveys have also been flown at less than 50 m elevation in selected areas. Flight lines are usually east–west, given the predominant NNW–SSE structural grain of the area. The plunge ($>60^\circ$) and limited strike length (<300 m) of some ore shoots highlight the need for a detailed aeromagnetic coverage (referred to in Figures 10–13 of Lindeman, 1984, with respect to gold exploration in the Murchison district). Aeromagnetic data for total magnetic intensity are capable of revealing subtle changes in the thickness and magnetization of the prospective ultramafic units and structural complexities on the prospective ultramafic–mafic contact. Nickel sulphide deposits may be characterized by areas of positive magnetic anomalies within the ultramafic pile (Figure 13.2).

The mafic–ultramafic contact is usually associated with a change in magnetic intensity, which is commonly greater than 100 nT. In addition to TMI data, horizontal and vertical derivative data are capable of further resolving details within the ultramafic ‘magnetic stratigraphy’. On a broader scale, magnetic lineaments revealed in regional aeromagnetic data have been spatially related to sites of nickel mineralization (O’Driscoll, 1981).

13.8.3 AIRBORNE RADIOMETRIC METHOD

13.8.3.1 Introduction

Radiometric surveys detect and map natural radioactive emanations, called gamma rays, from rocks and soils. Since gamma rays have the highest energy in comparison with alpha (α) and beta (β) particles and will penetrate several hundred feet through the atmosphere, only a few inches of Earth are sufficient to attenuate natural radiation by one-half. Therefore, the radioactive elements or their daughter products must occur in outcrop to be detected. All detectable gamma radiations from Earth materials come from the natural decay products of only three elements, namely, uranium (U), thorium (Th), and potassium (K).

Many man-made radioactive isotopes also emit gamma radiation. These man-made isotopes are produced by nuclear reactors or are the result of atomic weapons testing programmes. High-energy cosmic rays produced outside the Earth's atmosphere interact with the molecules of the atmosphere, the aircraft structure, and the detector itself to produce a variety of high-energy radiation. This cosmic ray component increases exponentially with the height above sea level.

In parallel with the magnetic method that is capable of detecting and mapping only magnetite (and occasionally pyrrhotite) in soils and rocks, the radiometric method is capable of detecting the presence of U (daughter products in the ^{238}U decay series), Th (daughter products in the ^{232}Th series), and K (potassium-40, which is 0.0118% of total potassium) at the surface of the ground. The basic purpose of radiometric surveys is to determine either the absolute or relative amounts of U, Th, and K in surface rocks and soils. The concentration and spatial distribution of radioactive elements such as uranium, thorium, and potassium from the upper layer (10–60 cm) of the Earth's surface having no vegetation are measured by radiometric surveys. The gamma ray detection indicates the abundance of these elements. Higher altitudes (above 300 m) may affect radiometric data negatively. Radiometrics cannot be used offshore or in permafrost areas since water attenuates the radiometric signal.

In order to recognize the incidence of usual radioactive elements such as uranium, thorium, and potassium, scintillometry (gamma ray method) is used. Multichannel spectrometers are used to detect individual radioelement abundances. The method has application in the exploration for uranium since it can be detected directly. Among the three radioelements, 'thorium' is the most immobile and resembles zirconium in behaviour. With the increase in alkalinity in rocks (as in felsic rocks), thorium content increases like uranium content. Radiometric surveys can cover large areas and are ideally suited for plane areas.

No other geophysical method, and probably no other remote sensing method, requires consideration of so many variables in order to reduce the observational data to a form that is useful for geological interpretation. Meteorological conditions, the topography of the survey area, the influence of the planet's cosmic environment, the height of the sensor above the ground, and the speed of the aircraft are just a few of the variables which affect radiometric measurements and which can bias an analysis unless dealt with very thoroughly.

13.8.3.2 General Applications of Radiometric Method

The radiometric method has a number of applications:

1. Obviously, it is most applicable to the direct search for uranium.
2. Since many rocks are naturally radioactive to characteristic degrees, the method is suitable to map geology by way of distinguishing different rock types. The methodology works more on the chemical response than the physical properties of the rock. In this way, different rock types and different phases may be distinguishable when they are not visually.
3. Since gamma rays cannot penetrate more than $\approx \frac{1}{2}$ m of rock, the natural response from rocks is, however, limited to this order of depth. The method is therefore able to map only the very near surface. In this way, it is useful in agriculture in mapping different soil types.

4. It can detect faults that act as conduits for radiation in the form of radon gas to escape to the surface.
5. Another important application of the method is its ability to determine the actual concentration of the radioactive elements and so 'assay' quantitatively rocks or drill core, in situ, if necessary. This can be quicker than having assays done in a geochemical laboratory and may be sufficient.
6. The progress of radioactive tracers used to define the rate of flow of liquids can be monitored using radiometric instruments.
7. Contamination in soil from leakage from nuclear reactors.
8. The presence of radioactive radon gas can be a precursor to earthquakes.

13.8.3.3 Instrumentation

The phosphor crystal and the photomultiplier tube (PMT) constitute the 'sensor', and another 'console' contains the high-voltage power supply for the PMT and electronics to register the voltage and display the 'count rate'. The whole sequence of one count lasts only a few microseconds. Time intervals for the accumulation of readings to obtain good statistics can be preset. The display is commonly digital, with LCDs or dials in older units. An audible output is also often provided. The crystal is commonly sodium iodide (NaI) that is 'doped' with thallium to improve its response. One type of instrument is a scintillometer, which measures the maximum count rate at specified energy levels. The highest energy level provides the 'total count'. A more advanced instrument is a spectrometer, which incorporates a pulse height analyzer to measure count rates in specified 'windows' or 'channels' of energy. Choosing the appropriate channels, the specific energy levels that characterize elements of interest can be identified. The elements most often measured are potassium, uranium, and thorium. A spectrometer that measures these three elements and total count is a four-channel type. A 256-channel spectrometer records the entire energy spectrum in much more detail but requires a bigger crystal volume to obtain sufficient counts in each channel. At least 8 L is needed for this, and since each litre weighs 4 kg, this 32 kg crystal requires some form of transport, such as a quad bike or a 4WD vehicle. A hand-portable crystal usually has a volume of (only) 1–2 L.

13.8.3.4 Statistics

Radioactive decay and, in this case, gamma radiation are a random process, and measurement over relatively short periods of time will not yield the true value of count rate (usually counts/sec, or cps, or counts/min, or cpm). It is possible, for example, because of the randomness, that no disintegrations occur for a long period of time and then a burst of them happens in a shorter time. Clearly, the count time is too short if it is less than the time during when no rays impinge on the detector. In fact, the probability that a given period will give the true rate decreases as the measurement time decreases.

The statistical 'percentage error' $E(\%)$ can be expressed as:

$$E(\%) = \frac{100}{\sqrt{n}}$$

where n is the number of counts collected in the time.

13.8.3.5 Site Geometry

Theoretically, and hence for calculations of element concentrations, it is assumed that measurements are taken on an infinite plane with uniform properties. In this case, the solid angle from the source to the detector is 2π . In practice, this means a flat, uniform area over a distance of about 5–10 m in all directions.

This requirement does not hold if measurements are taken on the crest of a hill, against a cliff, or in a valley. On the hill, the radiation from the surface will be less than for a flat surface, as less radiation will reach the detector, and against a cliff or in the valley, more will be detected.

To maintain a consistency of readings, care should be taken to keep the surface geometry as constant as possible, or if this is not possible, some corrections should be applied. In general, such corrections can only be approximate.

13.8.3.6 Detector Height

The height of the detector above a surface will influence the size of the area covered as the solid angle will vary with height. The higher the detector, the more the area measured. This, in turn, interacts with the choice of station interval for maximum coverage. A wider station interval requires a higher position of the detector. It is important for consistency and comparability of readings that the height should not vary on a survey grid. However, raising the height overall may be one way to increase counts if it means encompassing more of a background source.

13.8.3.7 Radiometric Assaying

The method can be used to derive actual element concentrations. Uranium and thorium concentrations are expressed in ppm, and potassium as a percentage. The values of concentration are derived by applying 'concentration coefficients' to the count readings. These are specific to each instrument and are deduced from measurements made with the instrument on 'pads' of known concentration of the three elements.

Estimates of uranium concentration are usually reported as 'equivalent uranium' (*eU*), as these estimates are based on the assumption of equilibrium conditions. Thorium is also usually reported as 'equivalent thorium' (*eTh*), although the thorium decay series is almost always in equilibrium. The energy spectrum of gamma ray is characteristic of the radioactive element it came from (Figure 13.3). Gamma ray can be measured in the ground or by a low-flying aircraft.

13.8.3.8 Background Corrections

The radiometric method is affected by meteorological variations. Principally, the radioactive radon gas that forms a background level can be blown about by wind, and it is also influenced by

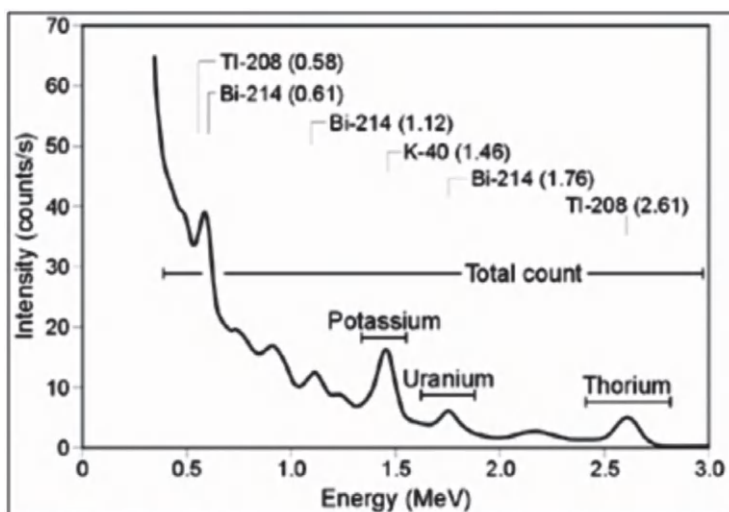


FIGURE 13.3 Gamma ray spectrum showing the position of the K, Th, U, and total count windows.

Source: International Atomic Energy Agency (IAEA, 2003).

barometric pressure and temperature. Rain can ‘wash’ the gas out of the air, and so the background may be reduced for a time after rainfall. Such events should be ‘commented’ upon. As radiation is strongly attenuated by water, the radiation from the soil and exposed rock can be greatly suppressed by standing water on the ground. This may be a temporary thing after rain. As many of these influences cannot be quantified, actual corrections for them can generally not be made. In any case, at least in ground surveying, the influences are minor. The particular conditions at the time of the survey should at least be noted.

13.8.3.9 Analysis and Interpretation

The gamma ray spectrometry method is widely used in diverse fields. Initially developed as a uranium exploration tool, the application of the method includes geological mapping, mineral exploration, soil mapping, and environmental radiation monitoring. In many situations, gamma ray spectrometry is probably more useful than any other single airborne geophysical or remote sensing technique in providing information directly interpretable in terms of surface geology.

U, K, and Th occur at high concentrations in fresh granite soil (white to violet) and in granitic soil, where clay layers are close to or exposed at the surface (yellow, pink, and white). The highest levels of U and Th (brightest blue/green) denote exposed ironstone gravel where potassium is low, and these colours fade out as the depth of overlying quartz sand increases. However, similarly muted U and Th signalling can emanate from duplex soil developed in highly weathered (K-deficient) substrate. Deep sandy duplexes and deep yellow and grey sands of the sandplain show up as brown to black, while deep granitic sands (fresh soil) with appreciable quantities of potassium feldspars show up as dark red.

Composite images provide a simultaneous display of up to three parameters on one image and facilitates the correlation and delineation of areas based on subtle differences in numerical values. The following combinations were developed by the USGS:

- The radioelement composite image combines the data of K (in red), eTh (in green), and eU (in blue).
- The potassium composite image combines the data of K (in red) with the ratios K/eTh (in green) and K/eU (in blue).
- The uranium composite image combines the data of eU (in red) with the ratios eU/eTh (in green) and eU/K (in blue).
- The thorium composite image combines the data of eTh (in red) with the ratios eTh/eU (in green) and eTh/K (in blue).
- The radioelement ratio composite image combines the data of the three radiometric ratios eU/eTh (in red), eU/K (in green), and eTh/K (in blue).

All processed radiometric data analyzed generated and correlated with geological and spectral interpretation. The radiometric spectral trends are corroborated with magnetic spectral strengths ([www: Geosiamservices.com](http://www.Geosiamservices.com); composite map of radiometric survey).

REFERENCES

- Bouligand, C., Glen, J.M. and Blakely, R.J., 2014. Distribution of buried hydrothermal alteration deduced from high resolution magnetic surveys in Yellowstone National Park. *Journal of Geophysical Research*, vol. 119, pp. 2595–2630.
- Brozena, J.M., Childers, V.A., Lawver, L.A., Forsberg, R., Faleide, J. and Eldholm, O., 2002. A new aerogeophysical study of the Eurasia Basin and Lomonosov Ridge: Implications for basin development. AGU Fall Meeting Abstracts, vol. 1, p. 1.
- Brozena, J.M., Childers, V.A., Lawver, L.A., Gahagan, L.M., Forsberg, R., Faleide, J.I. and Eldholm, O., 2003. New aerogeophysical study of the Eurasia Basin and Lomonosov Ridge: Implications for basin development. *Geology*, vol. 31, pp. 825–828.

- Camara, E.D.B. and Guimaraes, S.N.P., 2016. Magnetic airborne survey-geophysical flight. *Geoscientific Instrumentation, Methods and Data Systems*, vol. 5, pp. 181–192. Available at www.geosci-instrument-method-data-syst.net/5/181/2016/, DOI: 10.5194/gi-5-181-2016.
- Connors, G.T., 2013. Modelling the electromagnetic response of the helicopter-borne ZTEM system for vertical thin plate conductors buried below conductive overburden. In: *A Thesis Submitted to the Faculty of Science in Partial Fulfilment of the Requirements for the Degree of Bachelor of Science*. Ontario: Department of Earth Sciences Carleton University Ottawa, p. 41.
- Finn, C.A. and Morgan, L.A., 2002. High-resolution aeromagnetic mapping of volcanic terrain, Yellowstone National Park. *Journal of Volcanology and Geothermal Research*, vol. 115, no.1, pp. 207–231.
- Griffin, T.J., 1990. Geology of the granite-greenstone terrane of the Lake Lefroy and Cowan 1: 100000 sheets, Western Australia. Geological Survey of Western Australia, Report 32, p. 53.
- Guimaraes, S.N.P., Ravat, D. and Hamza, V.M., 2014. Combined use of the centroid and matched filtering spectral magnetic methods in determining thermomagnetic characteristics of the crust in the structural provinces of Central Brazil. *Tectonophysics*, pp. 624–625, 87–99.
- Heiskanen, W.A. and Moritz, H., 1967. *Physical Geodesy*. San Francisco: W.H. Freeman, p. 364.
- Hemant, K., Thébault, E., Mandeau, M., Ravat, D. and Maus, S., 2007. Magnetic anomaly map of the world: Merging satellite, airborne, marine and ground-based magnetic data sets. *Earth and Planetary Science Letters*, vol. 260, pp. 56–71.
- Hinze, W.J., Aiken, C., Brozena, J., Coakley, B., Dater, D., Flanagan, G. and Kucks, R., 2005. New standards for reducing gravity data: The North American gravity database. *Geophysics*, vol. 70, pp. J25–J32.
- Hohmann, G.W., 1987. Numerical modelling for electromagnetic methods of geophysics. In: Nabighian, M.S. (ed), *Electromagnetic Methods in Applied Geophysics—Theory*. vol. 1. Tulsa, OK: Society of Exploration Geophysicists, pp. 313–364.
- Hood, P. and Ward, S.H., 1969. Airborne geophysical methods. *Advances in Geophysics*, Elsevier, Netherlands, vol. 13, pp. 1–112.
- Kuzmin, P., Lo, B. and Morrison, E., 2005. Final Report on Modeling, Interpretation Methods and Field trials of an Existing Prototype AFMAG System, for Ontario Mineral Exploration Technology Program (OMET), Project No. P0102–007b. Ontario Geological Survey's Miscellaneous Release—Data series, Canada, Ontario Ministry of Natural Resources Publication MRD-167, p. 75.
- LaBrecque, J.L. and Ghidella, M.E., 1997. Bathymetry, depth to magnetic basement, and sediment thickness estimates from aerogeophysical data over the western Weddell Basin. *Journal of Geophysical Research, American Geophysical Union (AGU)*, United States, vol. 102, pp. 7929–7945.
- Labson, V.F., Becker, A., Morrison, H.F. and Conti, U., 1985. Geophysical exploration with audio frequency natural magnetic fields. *Geophysics*, vol. 50, pp. 656–664.
- Legault, J.M., Zhao, S. and Fitch, R., 2012. ZTEM airborne AFMAG survey results over low sulphidation epithermal gold-silver vein systems at Gold Springs, south eastern Nevada. 22ND ASEG International Geophysical Conference and Exhibition, Australian Society of Exploration Geophysicists, Extended Abstracts, p. 4.
- Lindeman, F.W., 1984. Geophysical case history of Water Tank Hill—Mt. Magnet, W.A. In: Doyle, H.A. (ed), *Geophysical Exploration for Precambrian Gold Deposits*. Geology Department & University Extension. Perth, Western Australia: The University of Western Australia, Publication no. 10, pp. 97–112.
- Morrison, D., 2004. Australian society of exploration geophysicists. *Preview Journal Issue*, vol. 109, p. 14.
- Nerem, R.S. and 19 co-authors, 1994. Gravity model development for TOPEX/POSEIDON: Joint gravity models 1 and 2. *Journal of Geophysical Research (Oceans)*, vol. 99, pp. 24421–24447.
- O'Driscoll, E.S.T., 1981. A broad-scale structural characteristic of major nickel sulphide deposits of Western Australia. *Economic Geology*, vol. 76, pp. 1364–1372.
- Rapp, R.H. and Pavlis, N.K., 1990. The development and analysis of geopotential coefficients models to spherical harmonic degree 360. *Journal of Geophysical Research*, vol. 95, pp. 21885–21911.
- Ravat, D., 1996. Analysis of the Euler method and its applicability in environmental magnetic investigations. *Journal of Environmental and Engineering Geophysics, United States*, vol. 1, pp. 229–238.
- Reynolds, J.M., 2011. *An Introduction to Applied and Environmental Geophysics*. 2nd edn. Chichester: John Wiley & Sons Ltd, p. 712.
- Salem, A. and Ravat, D., 2003. A combined analytic signal and Euler method (AN-EUL) for automatic interpretation of magnetic data. *Geophysics*, vol. 68, pp. 1952–1961.
- Slichter, L.B., 1955. Geophysics applied to prospecting for ores. *Economic Geology*, 50th Anniversary Volume, United States: El Paso, Texas, pp. 885–969.

- Swager, C., Griffin, T.J., Witt, W.K., Wyche, S., Ahmat, A.L., Hunter, W.M. and McGoldrick P.J., 1990. Geology of the archaean kalgoorlie terrane—an explanatory note. Perth, Western Australia: Geological Survey of Western Australia, Record 1990/12, p. 55.
- Telford, W.M., Geldart, L.P, Sheriff, R.E. and Keys, D.A., 1976. *Applied Geophysics*, Chapters 2 and 3. Cambridge: Cambridge University Press.
- Torge, W., 1989. *Gravimetry*. Berlin: de Gruyter, p. 465.
- Trench, A. and Williams, P.K., 2019. Nickel: Application of geophysics to nickel sulphide exploration in the Kambalda district, Western Australia. *ASEG Extended Abstracts*, vol. 1994, no. 1, pp. 169–180, DOI: 10.1071/ASEGSpec07_10.
- Vozoff, K., 1972. The magnetotelluric method in the exploration of sedimentary basins. *Geophysics*, vol. 37, pp. 98–141.
- Ward, S.H., 1959. AFMAG—airborne and ground. *Geophysics*, vol. 24, pp. 761–787.
- Ward, S.H., O'Brien, D.P., Parry, J.R. and McKnight, B.K., 1968. AFMAG-interpretation. *Geophysics*, vol. 33, no. 4, pp. 621–644.
- Ward, S.H., O'Donnell, J., Rivera, R., Ware, G.H. and Frazer, D.C., 1966. AFMAG- applications and limitations. *Geophysics*, vol. 31, no. 3, pp. 576–605.
- Yin, C.C., Zhang, B., Liu, Y.H. et al., 2015. Review on airborne EM technology and developments. *Chinese Journal of Geophysics*, China, vol. 58, no. 8, pp. 2637–2653 (in Chinese).

14 Remote Sensing Techniques, Global Positioning Systems (GPS), Geographic Information Systems (GIS)

14.1 REMOTE SENSING TECHNIQUES

Remote sensing is the acquisition of information about an object or phenomenon without making physical contact with the object and, thus, in contrast to on-site observation, especially on Earth. It is a technology used for the data acquisition of objects which are located at a distance or in remote regions and performs analysis for the interpretation (Sabins, 1999) of the physical attributes of the acquired data. Electromagnetic radiation (EMR) wavelengths ranging from visible to microwave regions are used for data acquisition in remote sensing (Gupta, 2003). Multispectral remote sensing imagery pixels have discrete spectral bands (Ting-ting and Fei, 2012), so it is not much helpful for the detection of minerals of the same classes (Pippi, 1989). Hyperspectral remote sensing is much helpful for the detection and mapping of different minerals of the same classes (Swayze et al., 1992). It provides spatial as well as spectral contiguous bands of the Earth surface for mineral mapping (Goetz et al., 1985). Spectral bands of a pixel give precise and accurate information about the objects. High-spatial-resolution (2–20 m) and high-spectral-resolution (5–10 m) data are provided by airborne sensors (Kruse, 2002; Kruse et al., 2003).

In current usage, the term *remote sensing* generally refers to the use of satellite or aircraft-based sensor technologies to detect and classify objects on Earth. It includes the surface and the atmosphere and oceans, based on propagated signals (e.g. electromagnetic radiation). It may be split into ‘active’ remote sensing (when a signal is emitted by a satellite or aircraft to the object and its reflection detected by the sensor) and ‘passive’ remote sensing (when the reflection of sunlight is detected by the sensor) (Lingyan et al., 2017; Schowengerdt, 2007).

The Geological Survey of India (GSI) is set to use an advanced remote sensing technology in 1986 developed by the National Aeronautics and Space Administration (NASA) to locate mineral deposits in the country. In a first-of-its-kind initiative, the organization has been installed with the Airborne Visible-Infrared Imaging Spectrometer–Next Generation (AVIRIS-NG) to explore mineral ores in 14 different locations. The sensor is an ‘engineering marvel’ and has proved effective for mapping surface mineralogy in different parts of the world. The AVIRIS-NG sensor was recently mounted on board the Indian Space Research Organisation (ISRO) aircraft to acquire hyperspectral images of mineral blocks in 14 locations, including Jhagadia and Ambaji in Gujarat, Tundi in Jharkhand, and Wajrakarur of Andhra Pradesh.

Multisensor airborne geophysical surveys are carried out by engaging with the Twin Otter Airborne Survey System (TOASS) and the Heliborne Geophysical Survey System (HGSS) as a significant aid in exploring mineral resources, particularly base metals. GSI uses digital image processing systems to utilize multispectral data as an aide to prime thrust areas, namely, mapping, mineral targeting, geoenvironmental hazard assessment, and geoengineering solutions. Hyperspectral remote sensing (Hyperion) is being used in GSI with the objective of utilizing the spectral response curve (spectral signature) of different alteration minerals like clay, iron minerals, and carbonates for their discrimination on the basis of their absorption features recorded in the spectra. It is used

in mineral mapping, particularly for hydrothermal alteration associated with a mineralized zone (www.gsi.gov.in; de Havilland Canada DHC-6 Twin Otter deployed by GSI in 1986 for aerogeophysical remote sensing survey; HAL Dhruv Helicopter deployed by GSI in 2014 for remote sensing heliborne geophysical surveys).

Remote sensing aerial surveys (RSAS) acquired the new heliborne geophysical survey system (HGSS) from PICO Envirotec Inc., Canada, in 2010. This system was installed and integrated into the newly purchased helicopter from Hindustan Aeronautics Ltd. The main geophysical sensors of this system included a magnetometer (Cs-3), a gamma ray spectrometer (GRS-10), a time domain electromagnetic system (PTHEM), and Gravimeter (GT-1A), along with various other ancillary equipment. The helicopter with all the installed sensors, christened as *Garuda Vasudha*, was ready for survey in the last week of February 2014 after obtaining all the requisite permissions from the Directorate General of Civil Aviation (DGCA). The Aldahalli area, Hassan district, Karnataka, was selected for heliborne surveys to compare and validate newly obtained survey data with earlier survey data obtained through the Bureau De Recherché Geologies' Et Meniere's (BRGM)/Compagnie Generale De Geophysique (CGG) of France in 1971–1972. The Chikmagalur airport was selected as the base of operation for these surveys.

BRGM had detected several time domain EM anomalies in the area which were inferred and categorized into good conductors (6 channel), moderate conductors (5 channel), and poor conductors (4 and 3 channel). After overcoming a series of teething troubles, the first successful test survey flight comprising P-THEM, spectrometric, and magnetic sensors was carried out on 3 April 2014. Four lines, viz., 1,700, 1,690, 1,680, and 1,670, were covered over the known conductors (associated with massive sulphides, subsequently proved by drilling) of the Aladahalli area, in an E–W direction, with a line spacing of 200 m and line length of 12 km each, maintaining the transmitting loop altitude of 50 m above the ground.

14.1.1 OVERVIEW

Passive sensors gather radiation that is emitted or reflected by an object or the surrounding areas. Reflected sunlight is the most common source of radiation measured by passive sensors. Examples of passive remote sensors include film photography, infrared, charge-coupled devices, and radiometers. Active collection, on the other hand, emits energy in order to scan objects and areas whereupon a sensor then detects and measures the radiation that is reflected or backscattered from the target. Radar and lidar are examples of active remote sensing, where the time delay between emission and return is measured, establishing the location, speed, and direction of an object.

Remote sensing makes it possible to collect data in dangerous or inaccessible areas. Remote sensing applications include monitoring deforestation in areas such as the Amazon Basin, glacial features in the Arctic and Antarctic regions, and depth sounding of coastal and ocean depths. Military collection during the Cold War made use of stand-off collection of data about dangerous border areas. Remote sensing also replaces costly and slow data collection on the ground, ensuring in the process that areas or objects are not disturbed.

Orbital platforms collect and transmit data from different parts of the electromagnetic spectrum, which, in conjunction with larger-scale aerial or ground-based sensing and analysis, provides researchers with enough information to monitor trends such as El Niño (El Niño is the warm phase of the El Niño–Southern Oscillation [ENSO] and is associated with a band of warm ocean water that develops in the central and east-central equatorial Pacific, between approximately the international date line and 120°W, including the area off the Pacific coast of South America) and other natural long- and short-term phenomena. Other uses include different areas of Earth sciences, such as natural resource management; agricultural fields, such as land usage and conservation (Howard et al., 2015); and national security and overhead, ground-based, and stand-off collection on border areas.

14.1.2 TYPES OF DATA ACQUISITION TECHNIQUES

The methods of imaging use visible, infrared, and thermal infrared electromagnetic spectrum, including satellite programs such as Landsat, SPOT, and IRS; high-resolution satellites; multispectral and hyperspectral techniques; microwave remote sensing; the types of radar systems; radar systems and laser systems; and drones used for observation of Earth surface.

India's space program began in the 1960s under the guidance of Dr Vikram Sarabhai. With the launch of IRS-1A in 1988, India started its own remote sensing program. To date, India has successfully launched tens of satellites, including high-spatial-resolution Cartosat satellites and radar imaging RISAT satellites. In that process, the polar satellite launch vehicle (PSLV) became one of the most successful satellite delivery systems. Apart from serving India's needs, the Indian remote sensing satellites also serve the requirements of its South Asian neighbours. Besides remotely sensing the Earth surface, India has also successfully planned and executed a moon mission (Chandrayaan) and a Mars mission (Mangalyaan) on the first attempt. India has also made considerable progress in human resource development in space technology and applications.

Mishra et al. (2017) reviewed the Indian remote sensing sensor system with respect to current and future prospects, where the authors argue that India has the world's largest constellation of remote sensing satellites in operation which is being extensively used for resource management and integrated planning for national development. They mentioned the series of Earth observation systems in both polar and geosynchronous orbits, starting from Bhaskara, the first experimental Earth observation satellite in 1979, to Cartosat-2E, successfully launched in 2017. They have reviewed active and passive sensors in both optical and microwave spectral regions, providing valuable data at different spatial resolutions for land, ocean, and atmospheric applications for the protection of the global environment, reduction of disaster losses, and achievement of sustainable development. The basis for multispectral collection and analysis is that of examined areas or objects that reflect or emit radiation that stand out from the surrounding areas. For a summary of major remote sensing satellite systems, see the overview in Table 14.1.

TABLE 14.1
Remote Sensing Satellite and Data Overview

Data Product	Provider	Orbital Height	Spatial Resolution	Swath Width
Landsat 5 TM	NASA/USGS	705 km	30 m, 120 m	185 km
		Sun-synchronous		

Data Use: oceanography, aerosols, bathymetry, vegetation types, peak vegetation, biomass content analysis, moisture analysis, thermal mapping, mineral deposit identification

Source:

1. NASA. Technical Details. Landsat Science. [Online] 18 September 2013. [Cited: 19 September 2013.] <http://landsat.gsfc.nasa.gov/?p=5081>.
2. USGS. Frequently Asked Questions about the Landsat Missions. USGS Landsat Missions. [Online] 29 August 2013. [Cited: 19 September 2013.] http://landsat.usgs.gov/tools_faq.php.

Landsat 7	NASA/USGS	705 km	15 m, 30 m, 60 m	183 km
ETM+		Sun-synchronous		

Data Use: oceanography, aerosols, bathymetry, vegetation types, peak vegetation, biomass content analysis, moisture analysis, thermal mapping, mineral deposit identification

Source:

1. NASA. Technical Details. Landsat Science. [Online] 18 September 2013. [Cited: 19 September 2013.] <http://landsat.gsfc.nasa.gov/?p=5081>.
2. USGS. Frequently Asked Questions about the Landsat Missions. USGS Landsat Missions. [Online] 29 August 2013. [Cited: 19 September 2013.] http://landsat.usgs.gov/tools_faq.php.

TABLE 14.1 (Continued)**Remote Sensing Satellite and Data Overview**

Data Product	Provider	Orbital Height	Spatial Resolution	Swath Width
Landsat 8	NASA/USGS	705 km	15 m, 30 m, 60 m,	185 km
OLI (B1-B9)		Sun-synchronous	100 m	
TIRS (B10-B11)				

Data Use: oceanography, aerosols, bathymetry, vegetation types, peak vegetation, biomass content analysis, moisture analysis, cloud cover analysis, thermal mapping, soil moisture estimation

Source:

1. NASA. Technical Details. Landsat Science. [Online] 18 September 2013. [Cited: 19 September 2013.] <http://landsat.gsfc.nasa.gov/?p=5081>.
2. USGS. Frequently Asked Questions about the Landsat Missions. USGS Landsat Missions. [Online] 29 August 2013. [Cited: 19 September 2013.] http://landsat.usgs.gov/tools_faq.php.
3. Landsat 8. [Online] 2013. [Cited: 19 September 2013.] <http://pubs.usgs.gov/fs/2013/3060/pdf/fs2013-3060.pdf>.

AVHRR/3	NOAA/USGS	830–870 km	1,090 m	833 km
Advanced Very-High-Resolution Radiometer		Sun-synchronous		

Data Use: cloud and surface mapping, land–water bounds, recognition of snow and ice, sea surface temperatures, and cloud cover analysis at night

Source:

1. Advanced Very High-Resolution Radiometer (AVHRR) Sensor Characteristics. Global Land 1-KM AVHRR Project. [Online] 23 June 2008. [Cited: 19 September 2013.] http://edc2.usgs.gov/1KM/avhrr_sensor.php.
2. NOAA. Advanced Very High Resolution Radiometer - AVHRR. NOAA Satellite Information System. [Online] 28 March 2013. [Cited: 19 September 2013.] <http://noaasis.noaa.gov/NOAASIS/ml/avhrr.html>.
3. United States Department of the Interior. AVHRR. National Atlas. [Online] 14 January 2013. [Cited: 19 September 2013.] http://nationalatlas.gov/articles/mapping/a_avhrr.html.

MODIS	NASA	705 km	250 m (B1–B2)	2,330 km
Moderate Resolution Imaging Spectroradiometer		Sun-synchronous	500 m (B3–B7) 1,000 m (B8–B36)	

Data Use: aerosols, land and cloud boundaries and properties, ocean biology, biogeochemistry, atm. water vapour, sea surface and atmospheric temperature, cloud analysis

Source:

1. NASA. Specifications. MODIS Web. [Online] [Cited: 19 September 2013.] <http://modis.gsfc.nasa.gov/about/specifications.php>.
2. USGS/NASA. MODIS Overview. LP DAAC. [Online] 19 August 2013. [Cited: 19 September 2013.] https://lpdaac.usgs.gov/products/modis_overview.

QuickBird	Digital Globe	482 km	65 cm B/W	16.8–18 km
		450 km	2.62 m RGBiR	
		Sun-synchronous	61 cm B/W	
			2.44 m RGBiR	

Data Use: mapping, change detection, planning (engineering, natural resources, urban, infrastructure), land use, EIA, tourism, military, crop management, environmental monitoring

Source:

1. Satellite Imaging Corporation. QuickBird Satellite Imagery and Satellite System Specifications. Satellite Imaging Corporation. [Online] 2013. [Cited: 19 September 2013.] www.satimagingcorp.com/satellite-sensors/quickbird.html.
2. Digital Globe. QuickBird - Data Sheet. Digital Globe. [Online] [Cited: 19 September 2013.] 'Archived copy' (PDF). Archived from the original (PDF) on 2013-06-19. Retrieved 2013-10-20.
3. GeoImage. Satellite Overview. GeoImage. [Online] 2012. [Cited: 19 September 2013.] www.geoimage.com.au/satellites/satellite-overview.

(Continued)

TABLE 14.1 (Continued)**Remote Sensing Satellite and Data Overview**

Data Product	Provider	Orbital Height	Spatial Resolution	Swath Width
GOES Geostationary	NOAA	35,790 km	1 km	Pacific Ocean,
Operational			4 km	Americas, and Atlantic
Environmental Satellite			8 km	(160°E to 20°W)
Ikonos	Digital Globe	681 km	80 cm B/W	11.3 km
		Sun-synchronous	3.2 m RBGiR	

Data use: mapping, change detection, planning (engineering, natural resources, urban, infrastructure), land use, EIA, tourism, military, crop management, environmental monitoring

Source:

1. GeoImage. Satellite Overview. GeoImage. [Online] 2012. [Cited: 19 September 2013.] www.geoimage.com.au/satellites/satellite-overview.
2. Digital Globe. IKONOS - Data Sheet. Digital Globe. [Online] June 2013. [Cited: 19 September 2013.] www.digitalglobe.com/sites/default/files/DG_IKONOS_DS.pdf.

SPOT	EADS Astrium	694 km	2.5 m	60 km
Satellite pour l		Sun-synchronous	5 m	
Observation de la Terre			10 m	
			20 m	

Data use: mapping, change detection, planning (engineering, natural resources, urban, infrastructure), land use, EIA, tourism, military, crop management, environmental monitoring

Source:

1. GeoImage. Satellite Overview. GeoImage. [Online] 2012. [Cited: 19 September 2013.] www.geoimage.com.au/satellites/satellite-overview.
2. EADS Astrium. Astrium GEO-Information Services SPOT International Price List. EADS Astrium. [Online] 2013. [Cited: 19 September 2013.] http://www2.astrium-geo.com/files/pmedia/public/r146_9_pricelist_spot_en_2012.pdf.
3. Technical Information about the SPOT Satellites. EADS Astrium. [Online] 2013. [Cited: 19 September 2013.] www.astrium-geo.com/en/191-spot-technical-information.

RADARSAT	Canadian Space Agency	RADARSAT1	8–100 m	45–500 km
		793–821 km	3 × 1 m–100 ×	18–500 m
	MDA	RADARSAT-2	100 m	5–500 km
		798 km	3 × 1 m–100 ×	
		RADARSAT	100 m	
		Constellation		
		5,927 km		

Data Use: environmental monitoring, ice monitoring, marine surveillance, disaster management, hydrology, mapping, geology, agriculture and forestry

Source:

1. Canadian Space Agency. Satellite Characteristics. Canadian Space Agency. [Online] 21 January 2011. [Cited: 19 September 2013.] www.asc-csa.gc.ca/eng/satellites/radarsat/radarsat-tableau.asp#r1.
2. MDA. Geospatial Services International. MDA Essential Information Solutions. [Online] 2012. [Cited: 19 September 2013.] <http://gs.mdacorporation.com>.

ASTER	NASA	705 km	15–90 m	60 km
Advanced Spaceborne	Government	Sun-synchronous		
Thermal Emission and	of Japan			
Reflectance Radiometer				

Data Use: vegetation, ecosystem dynamics, hazard and disaster monitoring, change detection, Earth science, land cover analysis

Source:

1. USGS/NASA. ASTER Data Access Policies. LP DAAC. [Online] 19 August 2013. [Cited: 19 September 2013.] https://lpdaac.usgs.gov/products/aster_policies.
2. ASTER Overview. LP DAAC. [Online] 19 August 2013. [Cited: 19 September 2013.] https://lpdaac.usgs.gov/products/aster_overview.
3. NASA. Aster. NASA Jet Propulsion Laboratory. [Online] 9 July 2004. [Cited: 19 September 2013.] <http://asterweb.jpl.nasa.gov/eos.asp>.

14.1.3 THRUST AREAS IN APPLICATIONS

Various application projects have been carried out using the present national as well as international Earth observation systems. Yet in the Indian context, the following areas are likely to receive more attention in the immediate future:

1. Management of natural resources to ensure sustainable increase in agricultural production
2. Study of the state of the environment, its monitoring, and the assessment of the impact of developmental actions
3. Update and generation of large-scale topographical maps
4. Exploration of marine and mineral resources
5. Operational meteorology and monitoring of land and oceanic processes to predict climatic changes.

14.1.4 DATA CHARACTERISTICS

The quality of remote sensing data consists of its spatial, spectral, radiometric, and temporal resolutions.

- *Spatial resolution.* The size of a pixel that is recorded in a raster image. Typically, pixels may correspond to square areas ranging in side length from 1 to 1,000 m (3.3 to 3,280.8 ft).
- *Spectral resolution.* The wavelength of the different frequency bands recorded. Usually, this is related to the number of frequency bands recorded by the platform. Current Landsat collection is that of seven bands, including several in the infrared spectrum, ranging from a spectral resolution of 0.7 to 2.1 μm . The Hyperion sensor on Earth observing 1 resolves 220 bands from 0.4 to 2.5 μm , with a spectral resolution of 0.10 to 0.11 μm per band.
- *Radiometric resolution.* The number of different intensities of radiation the sensor is able to distinguish. Typically, this ranges from 8 to 14 bits, corresponding to 256 levels of the greyscale and up to 16,384 intensities or 'shades' of colour in each band. It also depends on the instrument noise.
- *Temporal resolution.* The frequency of flyovers by the satellite or plane and is only relevant in time series studies or those requiring an averaged or mosaic image, as in deforesting monitoring. This was first used by the intelligence community, where repeated coverage revealed changes in infrastructure, the deployment of units, or the modification/introduction of equipment. Cloud cover over a given area or object makes it necessary to repeat the collection of said location.

14.1.5 DATA PROCESSING

In order to create sensor-based maps, most remote sensing systems expect to extrapolate sensor data in relation to a reference point, including distances between known points on the ground. This depends on the type of sensor used.

For example, in conventional photographs, distances are accurate in the centre of the image, with the distortion of measurements increasing the further you get from the centre. Another factor is, the platen against which the film is pressed can cause severe errors when photographs are used to measure ground distances. The step in which this problem is resolved is called georeferencing and involves computer-aided matching of points in the image (typically 30 or more points per image), which is extrapolated with the use of an established benchmark, 'warping' the image to produce accurate spatial data. As of the early 1990s, most satellite images are sold fully georeferenced. In addition, images may need to be radiometrically and atmospherically corrected.

14.1.5.1 Remote Sensing Software

A remote sensing software is a software application that processes remote sensing data. Remote sensing applications are similar to graphics software, but they enable generating geographic information from satellite and airborne sensor data. Remote sensing applications read specialized file formats that contain sensor image data, georeferencing information, and sensor metadata. Some of the more popular remote sensing file formats include GeoTIFF, NITF, JPEG 2000, ECW (file format), MrSID, HDF, and NetCDF.

14.1.5.2 Examples of Remote Sensing Software

1. Geomatica, PCI Geomatics
2. SAGA GIS (Open Source)
3. TNTmips GIS, MicroImages, USA
4. ERDAS IMAGINE
5. ENVI
6. Google Earth
7. GRASS GIS
8. OpenEV (Open Source)
9. Opticks (Open Source)
10. Orfeo Toolbox (Open Source)
11. RemoteView
12. SOCET SET
13. IDRISI
14. ECognition
15. ArcGIS

14.1.6 HYPERSPECTRAL VS MULTISPECTRAL REMOTE SENSING

Recent advances in remote sensing and geographic information have led the way to the development of hyperspectral sensors. Hyperspectral remote sensing, also known as imaging spectroscopy, is a relatively new technology that is currently being investigated by researchers and scientists with regard to the detection and identification of minerals, terrestrial vegetation, and man-made materials and backgrounds (Qingxi et al., 2014; the UAV-based hyperspectral imaging system). While multispectral imagery is useful to discriminate land surface features and landscape patterns, hyperspectral imagery allows for the identification and characterization of materials.

Imaging spectroscopy has been used in the laboratory by physicists and chemists for over 100 years for the identification of materials and their composition. Spectroscopy can be used to detect individual absorption features due to specific chemical bonds in a solid, liquid, or gas. Recently, with advancing technology, imaging spectroscopy has begun to focus on Earth. The concept of hyperspectral remote sensing began in the mid-80s and, up to this point, has been used most widely by geologists for the mapping of minerals. The actual detection of materials is dependent on the spectral coverage, spectral resolution, and signal-to-noise ratio of the spectrometer, the abundance of the material, and the strength of absorption features for that material in the wavelength region measured.

Hyperspectral remote sensing combines imaging and spectroscopy in a single system which often includes large datasets and requires new processing methods. Hyperspectral datasets are generally composed of about 100 to 200 spectral bands of relatively narrow bandwidths (5–10 nm), whereas multispectral datasets are usually composed of about 5 to 10 bands of relatively large bandwidths (70–400 nm). Hyperspectral imagery is typically collected (and represented) as a data cube with spatial information collected in the X–Y plane and spectral information represented in the Z-direction (Pu, 2017).

14.1.7 APPLICATIONS OF REMOTE SENSING IN GEOPHYSICAL EXPLORATION SURVEY AND CASE STUDY

The integration of different types of remote sensing data, along with ancillary data from different sources, is driving many new scientific investigations ranging from estimating forest biomass to mapping of Mars surface to find minerals. Remote sensing data is also helping to develop a better geographical information system (GIS), which, in turn, can be used for education, land management, natural resources management, environmental, and aeronautical applications. Government organizations are also opening their own portal to provide data to common citizens. Some of the major operational application themes in which India has extensively used remote sensing data are agriculture, forestry, biodiversity, groundwater targeting, water resources, land use, urban sprawl, geology, environment, coastal zone, marine resources, snow and glacier, disaster monitoring and mitigation, and infrastructure development. These are many applications which can take advantage of hyperspectral remote sensing.

14.1.7.1 Rheology of the Zagros Lithosphere from Postseismic Deformation of the 2017 M7.3 Kermanshah, Iraq, Earthquake

Postseismic deformation processes following the 2017 M7.3 Kermanshah, Iraq, earthquake using 2018–2020 Sentinel-1 InSAR time series data in order to visualize the fits of the postseismic models combining the ascending and descending displacement fields for the two time periods to the quasi-vertical and quasihorizontal displacements in east–west direction (Xiaoran et al., 2020). The north–south displacement cannot be calculated because only the right-looking InSAR data are available and because polar-orbiting sensors are not sensitive to north–south displacements. Simulations were generated for models with lower and higher viscosities in both the lower crust and the uppermost mantle using $\eta_{lc} = \eta_{um} = 10^{18}$ Pas for the low-viscosity model (model 2) and $\eta_{lc} = \eta_{um} = 10^{20}$ Pas for the high-viscosity model (model 3), respectively.

For time period 1, the low-viscosity model produces vertical displacements comparable to the observations, but the east–west displacements are different than observed (model 2, Xiaoran et al., 2020). This reinforces that afterslip was the dominating process during this period. For time period 2, the low- and high-viscosity models predict significantly more and less displacements than observed, respectively (models 2 and 3, Xiaoran et al., 2020), reinforcing that the viscosities are of the order of 10^{19} Pas (model 1, Xiaoran et al., 2020).

14.1.7.2 Mineral Mapping Techniques: A Case Study in South-Eastern Rajasthan

The high spectral resolution of hyperspectral sensors finds applications in various scientific domains like specific mineral identification and, hence, is capable of replacing the traditional techniques, such as multispectral remote sensing and field-based approach for mineral exploration (Jain et al., 2017).

Minerals identified from the imagery of airborne hyperspectral data are clay, iron ore, and few other. Minerals are mapped using the Spectral Angle Mapper (SAM) and Mixture Tuned Matched Filtering (MTMF) algorithms. Generated mineral maps are compared to know about the suitability of the algorithm for mineral identification. The MTMF algorithm mapped the minerals with higher accuracy, and mixing of one mineral class to another mineral class is also not done. The MTMF algorithm gives good results as compared to SAM algorithm. It maps the target minerals and rejects the erroneous positive pixels (Jain et al., 2017).

14.1.7.3 Integration of Remote Sensing, Gravity, and Geochemical Data for the Exploration of Cu Mineralization in Alwar Basin, Rajasthan, India

Remote sensing technologies have also evolved in the recent past to facilitate mineral exploration. Indicator minerals and a combination of lineament, geochemistry, and geophysical tools aim to find

Cu mineralization. The importance of remote sensing in mineral exploration to find potentially rich but unexplored zones has been established (Chatteraj et al., 2020).

Bouguer gravity data used in this study are point data extracted from WGM2012 Earth gravity model provided by the International Gravimetric Bureau (BGI). The interpolated Bouguer gravity map shows gravity values ranging from 57.11 to 97.30 mGal. Rock types within an area often contrast enough in density to cause gravity anomalies (WGM, 2012, BGI). For example, sedimentary rocks and their weathered derivatives that fill basins almost always have low densities due to the dominance of quartzo-feldspathic minerals and are characterized by gravity lows on anomaly maps. Mafic/basic rocks which contain high-density Fe-Mg minerals are often associated with gravity highs. These differences can be used to map large regions where rocks are inaccessible or concealed, to look for faults responsible for juxtaposing rocks of different densities or to infer structures such as basins, arches, and buried intrusions.

14.1.7.4 Salient Features of PTHEM (Time Domain Helicopterborne EM System) Test Survey Results of Preliminary Processing, Aladahalli Area, Hassan District, Karnataka, by the Geological Survey of India (GSI)

A preliminary processing was performed in the field using the PICO-supplied EM data extraction software, involving preliminary noise reduction and high-altitude correction for the removal of the transmitter effect and drift on the receiver signal, and some filtering, such as weighted average, median, etc., was applied to the TDEM dB/dt data (GSI e-News, 2014).

The resulting data was then windowed into different channels with varying lengths (in a logarithmic pattern) for proper visualization and analysis of the anomalies arising due to subsurface conductors. The figure in GSI e-News special issue (2014; RSAS Bangalore) shows the typical response over the good conductors on the western side, where the response is subdued in the early windows but persists strongly into the middle windows. The response over the poor conductors is stronger in the early windows but dies down rapidly in the middle and late windows. The response in the late windows is not clearly distinguishable from noise.

The different window data for all the lines were then gridded, and the resulting image was tiled one below the other to get a proper perspective of the quality of the conductors and their disposition in the subsurface. The tiled image of the grids for different window channels also shows that the response over the good conductors, though subdued in the early time channels, continues strongly in the middle to late time channels, corresponding to good conductors in the western part. But the response over the poor conductors is strong in the early time channels but does not continue in the middle to late time channels (GSI e-News Special Issue, 2014; RSAS Bangalore; Individual conductivity depth section; www.gsi.gov.in). Using the EMFLOW software, conductivity depth sections were prepared for the lines up to a depth of 250 m. Using the preceding conductivity depth sections, stripped conductivity depth images (spatial) were prepared at every 25 m level below the surface. The image slices are cascaded for visualizing the depth continuity (Figure 14.1).

14.1.7.5 Aerial Archaeology around Hornsburg in Lower Austria, Germany: Case Study

The terms *aerial* or *remotely sensed information* already indicate how aerial archaeology works: it uses the distant view. Archaeological sites show up on the ground surface by light–shadow contrasts (shadow marks), tonal differences in the soil (soil marks), or differences in height and colour of the cultivated cereal (crop marks). In that way, settlements, graveyards, and fortifications produce specific structures that can be identified easily from a high viewpoint. Aerial archaeology is one of the oldest prospection methods adopted first in England just after World War I where Roman and pre-Roman structures were found (Beazeley, 1919). Archaeologists have frequently used this technique since then before a site is excavated (Aitken, 1974; Binford, 1964). It is a very productive

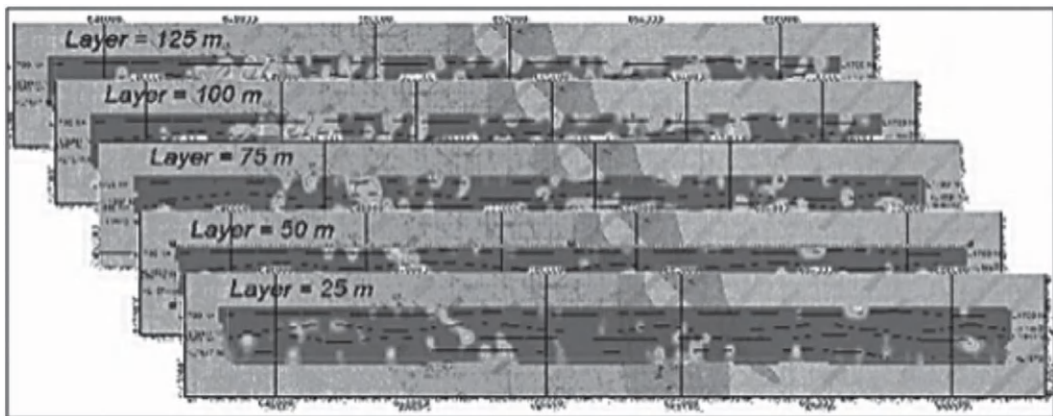


FIGURE 14.1 Conductivity depth image for every 25 m below the surface.

Source: GSI e-News (2014).

technique, and in relation to other methods, it is cheap, because it can cover quite a large area within a small number of flight hours—and any existing aerial photograph can be used for interpretation. A good example to illustrate the aerial photographs of the area around Hornsburg in Lower Austria is shown in the figure in Batayneh (2011) given by the Institute for Prehistory and Protohistory of the University of Vienna (1996).

Since the 1960s, airborne and spaceborne digital imagery has been used experimentally for archaeological prospection (Stringer and Cook, 1974; Lyons and Avery, 1977; Ebert, 1984) with only limited success. Berlin et al. (1977) have successfully used Landsat imagery to map areas in northern Arizona that had been cultivated and then abandoned almost 700 years ago. Perisset and Tabbagh (1981) have demonstrated that digital imagery in the infrared can be used in archaeological applications.

17.1.7.6 Locating Archaeological Sites in India: Case Study

Exploration and excavation are the two major components of any field archaeological research design and are mutually exclusive. Systematic ground surveys for locating archaeological sites, the extent of archaeological activity, the culture history of the settlement, etc. can be conducted prior to excavation and sometimes in lieu of excavation. Among numerous examples, mention should be made of the Vijayanagara Metropolitan Survey project at the World Heritage Site of Hampi, the Bellary and Kurnool Districts archaeological projects and survey of Palaeolithic landscapes in the Vindhyas and the Hunsig-Baichbal valleys, to name a few (Drewett, 2003). In the Indian context, aerial photographs from the sites of Tughlaqabad (Figure 14.2) and Sisupalgarh (Figure 14.3) are the best examples of the use of such techniques in the documentation of archaeological sites. It is interesting to note the significant changes in the landscape around Sisupalgarh between the images taken in 1948 and the recent Google Earth image. The aerial photograph of 1948 (Figure 14.3) shows open areas inside the fort and all around, whereas the recent Google Earth image clearly (Prabhakar and Korisettar, 2017) shows the extensive development and encroachment of areas close to the site.

However, an interesting feature in both these images is the presence of water bodies on the four corners of the fort, and another to the north-west of the fortification. Google Earth is the latest and simple tool for not only understanding various aspects related to terrain and geographical locations but also in locating archaeological sites and, in some, to observe more features that are not easily observable on the ground (Thakuria et al., 2013).



FIGURE 14.2 Aerial photograph of Tughlaqabad taken in 1945.

Source: Prabhakar and Korisettar (2017); Waddington (1946).



FIGURE 14.3 Aerial photograph of Sisupalgarh, Odisha, taken in 1948.

Source: Lal (1949); Prabhakar and Korisettar (2017).

14.2 GLOBAL POSITIONING SYSTEM (GPS)

14.2.1 INTRODUCTION

The Global Positioning System (GPS) is a satellite-based navigation and surveying system for the determination of precise position and time using radio signals from satellites, in real-time or in postprocessing mode. GPS is used all over the world for numerous navigational and positioning applications, including navigation on land, in air, and on sea; determining the precise coordinates of important geographical features as an essential input to mapping and Geographical Information System (GIS); along with its use for precise cadastral surveys, vehicle guidance in cities and on highways using GPS-GIS integrated systems, earthquake and landslide monitoring, etc. In India also, GPS is used for numerous applications in diverse fields, like aircraft and ship navigation, surveying, geodetic control networks, crustal deformation studies, cadastral surveys, creation of GIS databases, time service, etc., by various organizations.

The Navigation Satellite Timing and Ranging Global Positioning System (NAVSTAR GPS), developed by the US Department of Defence (DoD) to replace the TRANSIT Navy Navigation Satellite System (NNSS) by mid-1990s, is an all-weather high-accuracy radio navigation and positioning system which has revolutionized the fields of modern surveying, navigation, and mapping (US National Executive Committee for Space-Based Positioning, Navigation and Timing). For everyday surveying, GPS has become a highly competitive technique to the terrestrial surveying methods using theodolites and EDMs, whereas in geodetic fields, GPS is likely to replace most techniques currently in use for determining precise horizontal positions of points more than a few tens of kilometres apart. The GPS, which consists of 31 satellites in near-circular orbits at about 20,200 km altitude, now provides full coverage with signals from minimum four satellites available to the user at any place on Earth. By receiving signals transmitted by a minimum of four satellites simultaneously, the observer can determine his geometric position (latitude, longitude, and height), Coordinated Universal Time (UTC), and velocity vectors with higher accuracy, better economy, and less time compared to any other technique available today.

GPS is primarily a navigation system for real-time positioning. However, with the transformation from the ground-to-ground survey measurements to ground-to-space measurements made possibly by GPS, this technique overcomes the numerous limitations of terrestrial surveying methods, like the requirement of intervisibility of survey stations, dependability on weather, difficulties in night observations, etc. These advantages over the conventional techniques and the economy of operations make GPS the most promising surveying technique of the future. With the well-established high accuracy achievable with GPS in the positioning of points separated by a few hundreds of metres to hundreds of kilometres, this unique surveying technique has found important applications in diverse fields.

14.2.2 ACCURACIES WITH GPS AND COMPARISON WITH OTHER TECHNIQUES

GPS is the first positioning system to offer very high accuracy in most surveying and navigational applications at very low cost and with high efficiency. Within the span of a few years of its operation, GPS has truly revolutionized the field of surveying, with its potential to replace many conventional surveying techniques in use today. Accuracies now routinely achieved in the measurement of baseline lengths in relative mode, using high-precision geodetic instrumentation, with many hours of observations and scientific data processing, are as follows:

- 0.1–4 mm in local surveys (10–100 km baseline lengths)
- 4–10 mm in regional surveys (100–1,000 km baseline lengths)
- 1–2 cm in global surveys (1,000–10,000 km baseline lengths)

Such high accuracy standards make GPS suitable for various types of applications as compared to the limited range of applications of other positioning systems, like terrestrial surveying techniques,

inertial navigation system (INS), satellite laser ranging (SLR), very-long-baseline interferometry (VLBI), etc.

14.2.3 APPLICATIONS OF GPS

Due to the high accuracy, versatility, ease and economy of operation, and all-weather operation offered by GPS, it has found numerous applications in many fields, ranging from the millimetre-level high-precision geodesy to the several-metre-level navigational positioning. Some of these applications are:

- Establishment of high-precision zero-order Geodetic National Survey Control Network of GPS stations.
- Earth rotation and polar motion studies from GPS data.
- Estimating gravity anomalies using GPS.
- Marine geodesy: positioning of oceanic stations, buoys, etc.
- Earthquake monitoring: Crustal movements of the order of a few cm/year can be monitored using GPS method, thus making GPS the most suitable for monitoring continental drifts, neotectonic/seismotectonic movement, etc. The Department of Science and Technology of the Government of India and the Indian Space Research Organisation have initiated extensive GPS programmes for earthquake monitoring. In earthquake-prone areas, such as the Pacific Rim, GPS is playing an increasingly prominent role in helping scientists anticipate earthquakes. Using the precise position information provided by GPS, scientists can study how strain builds up slowly over time in an attempt to characterize, and in the future, perhaps anticipate, earthquakes.
- Geophysical positioning, mineral exploration, and mining.
- Survey control for topographical and cadastral surveys.
- Offshore positioning: shipping, offshore platforms, fishing boats, etc.
- Spacecraft tracking: vector separation between GPS satellites and any other satellites can be monitored by GPS, for example, pinpointing the location of Landsat, etc.
- General aircraft navigation, approach to runways, navigation/positioning in remote areas like deserts, dense jungles, shaded areas of microwave, precise sea navigation, approach to harbours, etc. It is expected that in the 1990s, most civilian aircrafts, ships, and boats will be fitted with GPS equipment, and even hikers, boat and car owners, and truck drivers will be using it extensively.
- Military: improved weapon delivery accuracies, that is, for missiles, etc., for ranging in artillery; navigation for Army, Navy, and Air Force, thus affecting ultimate saving of up to 1 billion dollars annually on navigation in the USA.
- Scientific applications, like studies related to the ionosphere and troposphere, glaciology, etc.

14.2.4 GPS IN INDIA

Developing countries like India are also entering the GPS era in a big way. Many scientific government organizations, including Survey of India; Indian Institute of Geomagnetism, Mumbai; National Geophysical Research Institute, Hyderabad; National Informatic Centre; Wadia Institute of Himalayan Geology, Dehradun; Centre for Mathematical Modelling and Computer Simulation, Bangalore; National Remote Sensing Agency; and many educational institutes and universities, like the Indian Institute of Technology, Bombay and Kanpur; University of Roorkee; Anna University, Chennai; Indian Institute of Remote Sensing, Dehradun; etc., are involved in GPS-related activities for various applications. Training courses in GPS are also being introduced at different institutes and universities. Many other nongovernment scientific, educational, and commercial organizations

have also procured GPS equipment for different applications, including crustal movement studies for earthquake monitoring, national highways project, city guide maps, GIS, etc. The Armed Forces are also in the process of adopting this new technology. The problems posed by the selective availability, antispoofing, and transformation of the GPS coordinates from WGS-84 to the Everest datum are being investigated in order to ensure an economic, efficient, and effective implementation of GPS technology in India. The need for establishing a national GPS centre to coordinate the activities of the various organizations in this field and to explore the feasibility of establishing a GPS tracking station network in India to obtain precise positioning of the satellites is being felt increasingly.

14.2.4.1 Global Positioning System (GPS) Data in India

In India, crustal deformation studies using global navigation satellite system (GNSS) (mostly GPS) started with the establishment of a permanent international GNSS service (IGS) station at the Indian Institute of Science (IISc), Bangalore, in the early 1990s. This was followed by the development of a national programme on crustal deformation studies in 1997–1998. Several campaign mode observations of plate motion and crustal deformation using GPS were initiated. In 1997, another IGS GPS station was installed at NGRI Hyderabad. By the beginning of this century, the programme had expanded extensively. Presently, there are about 100 permanent or semipermanent GPS stations in India maintained by different organizations, such as the National Geophysical Research Institute (NGRI), Survey of India (SOI), the Centre for Mathematical Modelling and Computer Simulation (CMMACS), etc. Several campaign mode studies in the seismically active regions have been taken up through periodical GPS observations for a few days. All permanent stations are installed on elevated ground, away from trees and buildings, to avoid obstructions in tracking the GPS satellites. Generally, a hard rock site is preferred so that actual ground motion can be recorded. An RCC pillar of height 6–8 ft, reinforced with steel rods, is erected (Figure 14.4a) on the hard rock. In case the hard rock is not exposed, the pillar is erected right from the foundation which either reaches the firm ground or is at least six feet below the ground surface.

The semipermanent stations are also made on firm base (Figure 14.4b). The locations of the very-small-aperture terminal (VSAT)–connected permanent stations operated by different organizations are plotted in the figure in Tiwari et al. (2014). All the permanent stations acquire GPS data at 30 sec sampling interval, and a daily file (called as the RINEX observation file) is generated. This



FIGURE 14.4 (a) A typical permanent GPS station at Port Blair, Andaman. (b) A typical semipermanent GPS station at Yamunotri in the Himalayan region.

Source: Tiwari et al. (2014).

file is used to compute the coordinates of the site, which could be accurate at ~ 3 mm level. The time series showing the daily variation of these coordinates provides an estimate of site motion, which, when compared with the plate motion, provides estimate for the crustal deformation in that region. Many of these stations are now connected through VSAT, and all the data are archived at the Indian National Centre for Ocean Information Services (INCOIS), Hyderabad, and the Indian Meteorological Department (IMD), New Delhi, data centres. There is a data sharing policy, and these data can be made available to any researcher/officials on request from INCOIS (www.isgn.gov.in). These data are now extensively used in earthquake studies besides atmospheric studies, ionospheric studies, control surveys, aircraft navigation, etc. (Banerjee et al., 2008; Mahesh et al., 2012; Gahalaut et al., 2013).

14.2.5 APPLICATIONS OF GPS SURVEY IN GEOPHYSICAL EXPLORATION SURVEY AND CASE STUDY

Geodetic measurements obtained with the Global Positioning System (GPS) are increasingly more widely applied in geophysical studies. Geophysical investigations employing GPS to measure coseismic, postseismic, and interseismic deformation; plate motion and crustal deformation at plate boundaries; volcano deformation; glacial isostatic adjustment and sea level change related to geodynamic and earthquake studies—the use of GPS determinations for the modelling of this study of geophysical phenomena.

14.2.5.1 Andaman and Nicobar Islands: GPS Survey by GSI, FSP 2005–2008

The Geological Survey of India has carried out a GPS survey in the Andaman and Nicobar Islands for the detection and quantification of postearthquake ground deformation during crustal adjustments after tsunamis on December 2004, under FSP 2005–2008.

The Geological Survey of India (GSI) launched 22 permanent Global Positioning System (GPS) stations across India to identify seismically hazardous zones and encourage mapping activities. The 22 GPS-geodetic (pertaining to *geodesy*, the science of Earth measurement) observatories are located in Kolkata, Thiruvananthapuram, Jaipur, Pune, Dehradun, Chennai, Jabalpur, Bhubaneswar, Patna, Raipur, Bhopal, Chandigarh, Gandhinagar Vishakhapatnam, Agartala, Itanagar, Mangan, Jammu, Lucknow, Nagpur, Shillong, and Little Andaman. The GSI has taken a stride to establish and maintain a network of continuously operating 35 permanent GPS stations in a phased manner. The new 13 stations will be located in Aizawl, Faridabad, Uttarkashi, Pithoragarh, Cooch Behar, Zavar, North Andaman, Middle Andaman, South Andaman, Ranchi, Mangalore, Imphal, and Chitradurga. These stations are meant to delineate high-strain zones for earthquake probability, determine a seismic motion on faults that may lead to a rupture, and produce thematic maps with high positional accuracy.

14.2.5.2 Coseismic Crustal Deformation from the 1994 Hokkaido–Toho–Oki Earthquake Monitored by a Nationwide Continuous GPS Array in Japan

Coseismic displacements (Segall and Davis, 1997; Tsuji et al., 1995) have been determined with GPS for at least 15 earthquakes (Table 14.2) in the Kurile Islands, Japan, earthquake on October 1994.

14.2.5.3 Coseismic Displacements of the 1994 Northridge, California, Earthquake

The M6.7 January 1994 Northridge earthquake was possibly the best geodetically monitored earthquake yet to occur. Vector displacements were determined at 66 GPS sites (Hudnut et al., 1996). Hudnut et al. (1996) used the random cost method and found that the best-fitting rectangular dislocation does not pass through the principal concentration of aftershocks but is displaced 1–2 km into the hanging wall (Hudnut et al., 1996).

TABLE 14.2**Recent Earthquakes for Which Coseismic Displacements Have Been Determined by GPS**

Earthquake	Magnitude (M)	References
1987 Superstition Hills, California	6.2	Larsen et al. (1992)
1989 Loma Prieta, California	7.1	Lisowski et al. (1990), Williams et al. (1993)
1989 Kalapana, Hawaii	6.1	Dvorak (1994)
1991 Valle de la Estrella, Costa Rica	7.7	Lundgren et al. (1993)
1992 Cape Mendocino, California	7.0	Murray et al. (1996)
1992 Joshua Tree, California	6.1	Bennett et al. (1995)
1992 Landers, California	7.3	Blewitt et al. (1993), Bock et al. (1993), Hudnut et al. (1994), Freymueller et al. (1994)
1993 Guam	7.8	Beavan (1993)
1994 Northridge, California	6.7	Hudnut et al. (1996) Shen et al. (1996)
1994 Arthurs Pass, New Zealand	6.7	Árnadóttir et al. (1995)
1994 Hokkaido–Toho–Oki, Japan	8.1	Tsuji et al. (1995)
1994 Sanriku–Haruka–Oki, Japan	7.5	Miyazaki et al. (1996)
1995 Hyogo-ken Nanbu (Kobe), Japan	7.2	Tabei et al. (1996)
1995 Anatofagasta, Chile	8.1	Ruegg et al. (1996)
1995 Jalisco, Mexico	8.0	Melbourne et al. (1996)

Source: After Segall and Devis (1997).

14.2.5.4 GPS Applications for Geodynamics and Earthquake Studies: Sanriku–Haruka–Oki Earthquake, Japan

A striking example of postseismic is offered by the Japanese national GPS array, which recorded a transient following the Sanriku–Haruka–Oki earthquake (Segall and Davis, 1978). Postseismic deformation followed the M7.5 Sanriku–Haruka–Oki, Japan, earthquake. Shown are estimates of the east component of displacement of station Kuji on the east coast of Honshu, Japan, determined every 75 min using data from the Japanese permanent GPS network. Note that the postseismic displacement is 25–40% of the coseismic signal. The north component shows a considerably smaller (<2 cm) transient.

14.2.5.5 Plate Motion and Crustal Deformation Estimated with Geodetic Data from the GPS (Pacific, North America, Eurasia, Australia, Africa, and South America)

Plate motion models based entirely on GPS data, which are dominantly from permanent GPS receivers at IGS sites (Segall and Davis, 1997; Argus and Heflin, 1995). Argus and Heflin (1995), estimated the motion of six major plates (Pacific, North America, Eurasia, Australia, Africa, and South America) using IGS data for 43 globally distributed stations. The measured angular velocities agree with the NUVEL-1A (DeMets et al., 1990, 1994) model within uncertainties for all plate pairs. These results are consistent with previous studies that have shown that the relative motion of plate interiors does not change measurably over time scales of years to millions of years. Larson and Frey Mueller (1995) used three years of IGS data from seven sites to estimate the motion of the Australian, Pacific, and Antarctic plates. Their study represents the first geodetic determination of Antarctic plate motion. The estimated rates agree with the NUVEL

no-net-rotation model of absolute plate motion to within error measurements. GPS holds great promise in the study of volcano deformation, which can be both episodic and rapid. Measurable deformation precedes some eruptions, which suggests the utility of GPS measurements in predicting volcanic hazards. Indeed, some dramatic GPS data have already been obtained related to volcanic processes.

14.2.5.6 Rapid Deformation of the South Flank of Kilauea Volcano, Hawaii, United States of America

Rapid quasisteady motions have been mapped on the Kilauea volcano, Hawaii, using GPS (Owen et al., 1995). The south flank of Kilauea, the area south of Kilauea's two rift zones, has generated large earthquakes, including a M7.2 event in 1975. These quakes occurred on a buried subhorizontal fault (with the upper plate moving seaward) near the base of the volcanic pile at the contact with the ancient seafloor. GPS data collected between 1990 and 1993 show that the region north of the rift zones was not actively deforming, while the central south flank displaced seaward at rates of up to 10 cm/year (Segall and Davis, 1997; Owen et al., 1995).

14.2.5.7 BIFROST Project (1996): Europe

GPS-derived determinations of deformation, particularly uplift, in areas of ongoing postglacial rebound will thus add invaluable data for testing models of mantle viscosity and ice load history. The first significant vertical deformation rates to be determined by GPS have only recently become available (BIFROST Project 1996). The figure in Segall and Davis (1997; Mitrovica et al., 1994) shows GPS estimates of vertical motion in Fennoscandia obtained from 2.5 years of observations. As three-dimensional crustal velocity estimates obtained from this continuing project become more accurate, they will be used to refine models of mantle viscosity and ice history.

14.2.5.8 Gravitational Sliding of the Mt Etna (East Coast of Sicily) Massif along a Sloping Basement: Case Study by GPS Survey

Mt Etna volcano dilates horizontally between its major flank eruptions. This outward expansion has been attributed to two processes: (1) gravitational spreading of its tall edifice and (2) inflation of a magma chamber within the volcano prior to eruption (Borgia et al., 2000; Lundgren et al., 2004; Neri et al., 2004; Obrizzo et al., 2004; Bonaccorso et al., 2006, 2013). However, it has long been noted that intereruptive horizontal displacement vectors measured from repeated GPS readings radiate not from the summit but from an area WNW of it, leading to suggestions that the position of the inflating magma chamber is offset from the summit by up to a few kilometres, or that the magma chamber is tall and slopes in this direction (Nunnari and Puglisi, 1994; Bonaccorso et al., 2011). This, in turn, implies that magma chamber inflation is the dominant process causing Etna's dilation, since displacement vectors caused by gravitational spreading of a conical volcano should be radial to the summit of the edifice. Another peculiarity is the lack of symmetry in vector lengths, which are consistently much longer on the ESE side of the volcano. This has been interpreted with numerical and analytical models, in which the symmetrical radial movement of the inflating magma chamber is distorted by two large rectangular subhorizontal dislocation planes dipping in different directions that underlie the southern and eastern flanks in 1994–1995 (Bonforte and Puglisi, 2003), or by a single larger dislocation plane beneath the lower eastern flank (Lundgren et al., 2003; Bonaccorso et al., 2006, 2013), or by a drag force due to subsidence of the continental margin on that side (Bonforte et al., 2011).

Murray et al. (2018) conducted recent GPS measurements which comprise over 100 benchmarks on Etna and the surroundings (Murray et al., 2018) and measured once a year with Leica system 530 GPS kits. The times of surveys referred to in this study are August 2001, September 2002, September 2003, October 2004, September 2005, September 2006, October 2007, September 2008, and September 2012. Results of the measurements are presented in the horizontal displacement

vector maps shown in the figure in Murray et al. (2018) for the intereruptive (major flank eruptions) periods (2001, 2002, 2005, 2007, 2008, and 2012). All these eruptions caused metre-scale displacements consequent upon dyke injection at the start of the eruption. The vectors in the figure in Murray et al. (2018) show that dilation occurs in each intereruptive period and that it is off-centre and asymmetric.

The horizontal displacements that occurred during flank eruptions are illustrated by the series of graphs shown in the figure in Murray et al. (2018). A dilating volcano that expands perfectly radially will produce horizontal displacement vectors that are axisymmetric and radial to the centre of displacement (Murray et al., 2018). In the figure in Murray et al. (2018), a map of vector lengths expected from a Mogi point source at 10 km depth is shown to have raised the surface by 20 cm (radial black arrows). In the figure in Murray et al. (2018), a sliding vector of 6 cm with an azimuth of 120° (unidirectional grey arrows), roughly corresponding to the main direction of basement slope at Etna (Neri and Rossi, 2002; Branca and Ferrara, 2013), is added to each radial vector. In the figure in Murray et al. (2018), the vectors resulting from adding the sliding vectors are shown. The choice of 6 cm for the sliding vector is intended to be purely illustrative. A glance at the displacement vectors measured on Etna (Murray et al., 2018), particularly between 2008 and 2012 (Murray et al., 2018), clearly shows similar characteristics. Once the sliding vector has been derived for each point and subtracted from the observed displacement vector, the deformation without sliding can be portrayed, and any other deviations from radial movement highlighted. This has been done in the figure in Murray et al. (2018) for the period 2008–2012.

It can be clearly seen that the vector lengths east and west of the volcano are much more balanced, and that most of the vectors radiate from the summit, rather than from an area WNW of it. The figure in Murray et al. (2018) shows how closely the vectors correspond to radial movement: a linear fit of the direction from the summit (angle u) against the azimuth of corrected displacement vector gives a coefficient of determination r^2 of 0.94 (Murray et al., 2018). Looking at those vectors that deviate from radial, all of them are close to areas of local faulting (Borgia et al., 1992; Allard et al., 2006; Obrizzo et al., 2001; Walter et al., 2005; Alparone et al., 2012). Benchmarks near the Pernicana, Ragalna, and Mascalucia fault complexes (Murray et al., 2018) all deviate from radial by more than 20° . This is not surprising in the case of the Pernicana fault, which underwent acceleration in 2010, right in the middle of the 2008–2012 period (Guglielmino et al., 2011).

Murray et al. (2018) also looked at the effect of sliding at a volcano experiencing gravitational spreading alone, by re-examining the analogue model shown in the figure in Murray et al. (2018) (Wooller et al., 2004). The figure in Murray et al. (2018) shows the setup of the analogue model (Wooller et al., 2004), and below it (Murray et al., 2018), a section through Mt Etna showing the Etna cone and lavas and the weak and ductile Quaternary sediments beneath. The entire setup is tilted by 1° to represent the effects of a sloping basement, and the figure in Murray et al. (2018) shows the vectors of movement after 15 min. Because the slope of the Etna basement is roughly 120° azimuth, the figure in Murray et al. (2018) should be rotated clockwise by 30° to compare with the Etna results in the figure in Murray et al. (2018). The 2004 eruption was noteworthy for the generally small amounts of deformation, behaving like a summit eruption in that respect, so can be regarded as similar to an intereruptive period. By subtraction of the sliding vector from Bonforte et al.'s (2008) published data (Murray et al., 2018), it is clear that horizontal contraction has indeed occurred all around the summit, with vectors pointing towards the summit, the eruption site, and the area of subsidence on the north flank. Like the period 2008–2012, also seen is clear displacement associated with the well-known areas of local faulting, particularly the Pernicana fault, where a small left-lateral displacement averaging 4 cm has taken place, though this movement diminishes with distance from the fault.

The sloping model basement used in the laboratory analogue model is planar, in contrast to the Etna basement surface, which is an almost-horizontal plateau surface beneath the NW flanks, with a 17 km wide horseshoe-shaped depression scooped out of the eastern and south-eastern sector

(Neri and Rossi, 2002; Norini and Acocella, 2011; Branca and Ferrara, 2013) that reaches a depth of about 400 m and includes the summit craters, which lie about 1 km inside its north-eastern edge. The overall slope of the basement is towards the ESE, and the figure in Murray et al. (2018) shows a WNW–ESE section through the volcano showing the basement topography, compared to the model planar substratum (Murray et al., 2018).

Regarding the cause of the component of radial expansion between eruptions, two different models are used. One simulates a magmatic inflation on a sliding basement (Murray et al., 2018) in which the component of radial expansion is the direct result of magma pressure, and the other gravitational spreading on a shallow slope (Murray et al., 2018) in which radial expansion is caused by the volcano deforming under its own weight (Borgia et al., 1992; Borgia, 1994). Both models reproduce the main features of Etna's horizontal deformation; it remains to be seen which of the two processes is dominant. Both models reproduce the main features of Etna's horizontal deformation; it remains to be seen which of the two processes is dominant. In this study, only horizontal movements were considered; adding the vertical component of movements could help discriminate the two models. The present model demonstrates that if inflation of a magma chamber is causing the dilation of Etna, then the chamber is directly beneath the summit, rather than WNW of it, as suggested by previous authors (Nunnari and Puglisi, 1994; Puglisi et al., 2004; Bonaccorso et al., 2011). This is the first time that current persistent basement sliding of the entire edifice has been detected and measured on an active volcano. It is concluded that the entire edifice of Mt Etna is sliding downslope towards the Mediterranean Sea at an average rate of 14 mm per year. The sliding is lubricated by the weak sediments beneath Etna and effected by the slope of the basement. Murray et al. (2018) proposed that the complex depression in the basement/volcano interface beneath the ESE flanks of the volcano (Murray et al., 2018) is the result of the gradual sinking of the volcanic pile into the weaker basement.

These results are important because there is strong geological evidence that volcanoes that have slid downslope in this manner have a propensity to experience very large catastrophic sector collapse on the downslope side later in their history, as at Socompa, Chile (Wooller et al., 2004), and at Colima, Mexico, which is situated on the steep slopes of the larger, older Nevado da Colima (Cortes et al., 2010). Such events are rare, occurring about four times a century worldwide (Siebert, 1992), but have devastating consequences, so the possibility of future occurrences needs to be taken seriously, both at Etna and in other volcanoes.

14.3 GEOGRAPHIC INFORMATION SYSTEMS (GIS)

14.3.1 INTRODUCTION

Geophysics is a kind of natural science for studying the Earth using gravity, magnetic, electrical, and seismic methods. It has wide application areas, especially in geophysical prospecting. The geographic information system (GIS) allows us to visualize, analyze, and interpret data to understand relationships, patterns, and trends, and it has a strong power of mapping. GIS has been used in various fields, such as crime mapping, landscape architecture, and regional and community planning, and it benefits organizations of all sizes and in almost every industry. Among them, the application in geophysics is a comparatively new growing respect. Although not so mature as applications in other fields, GIS has played a significant role in nearly all procedures of *geophysics studies*, including data collection, data storage and management, modelling, and interpretation, since the inception of integration of GIS and geophysics, which open up an era in GIS history.

Geophysical investigations usually involve huge volumes of diverse data out in the field and compute intensive processing algorithms for data manipulation back in the laboratory. When collecting data, the first step should be reconnaissance in field, and then to design reasonable and efficient sampling sites and measuring lines. However, conducting initial survey information by walking around the whole section of study area could cost a lot of money and time, especially when

the operation region contains rough areas, such as deserts, plateaus, and forests with steep slopes and valleys. Under this condition, GIS proves to be very effective. Researchers could integrate the coordinate data of sampling sites and the digital topographic map so that they can figure out the surface features that are passed through by measuring lines. As a result, initial sampling points could be easily shifted to better locations if they are too difficult or even impossible to access or obstructive to high-value economic crops. For instance, a project done in Ethiopia for oil and gas exploration using GIS provided a great proof of application of GIS in this aspect. Investigators there successfully avoided the valley in the middle of the seismic measuring net by using *ArcGIS*. After measuring lines planning, field geophysical data collection takes place.

When geophysical data has been acquired in the field, follows up is the problem of storing and managing the huge amount of various kinds of data. With the constant development of computer technology, it becomes possible to build up a database for geophysical investigation. Putting in the basic data is a relatively simple process. Software like FOXBASE and FOXPRO can be put into use directly beyond the function of database build-up which comes with the GIS software. Owing to the complexity of a geophysical study inclusive of a huge mountain of different diversities of data, it would be better to use large-scale relational database and combine it with GIS organically. During the constructing process, researchers can put in the data of date, coordinates of sampling points, gravity, magnetism, electricity, seism, and so on. Besides, building up a database for different methods separately on the basis of actual requirement is also an excellent approach. For example, gravimetric database can incorporate data of coordinates of measuring points, their elevation, observed gravity, topographic correction, free space gravity anomaly, Bouguer gravity anomaly, isostatic gravity anomaly, and so forth. In the database, the coordinates of measuring points are the most fundamental components for data processing and comprehensive interpretation, because other kinds of data are connected with each other, relying on these coordinate data. As technology advances and geophysical study go on, the editions of data modification, addition, deletion, and displacement with new data could be achieved quite easily using the function of data management of GIS. Furthermore, GIS has data retrieval capability, with which investigators can query and retrieve data in the entire database with ease.

Additionally, GIS performs as a powerful tool when analyzes compositely. More specifically, on the foundation of database, we can do the secondary development of database using GIS, by which we can reset or splice data and maps flexibly according to practical needs, and then yield various research results. First of all, GIS allows us to create diverse basic maps, such as measuring points distribution map, Bouguer gravity anomaly map, survey lines distribution map, etc., making use of its projection, network, and equivalent functions, based on the information of measuring points' coordinates. An important aspect of GIS techniques, networking data has immensely positive impact on geophysical data, for the two-dimension information of geophysical field is much more desired by geoscientists. Secondly, GIS enables humans to synthetically analyze data from diverse sources and spectrums, which is improbable to realize in traditional work. We can easily take advantage of the GIS overlay method to merge different features into a new coverage and combine disparate coverages into a new coverage in accordance with actual requirements for further study and interpretation. Therefore, using GIS in geophysics can efficiently reduce a good deal of repetitive work and allows different types of information to be used in concert with each other.

Ultimately, GIS helps establish three-dimensional models of the subsurface of the Earth for geophysical study. Modern interpretations in geophysics require an interdisciplinary knowledge, integration, and consideration of state-of-the-art information from comprehensive databases. Towards this end, a combination of disparate geophysical surveys employing gravity, magnetics, seismic, and geoelectric, together with geological, petrological, mineralogical studies, and results from remote sensing, provides new insights into the structures and tectonic evolution of natural deposits and the crustal underground. There is no doubt that the interdisciplinary approach and means are essential for numerical modelling of these structures and the process acting on it. Due to a variety of

geophysical methods, an increasing number of models have been yielded in most of geophysical data interpretation. However, only a small quantity of them is in three dimension, while most of them are still in two dimension, and some even in one dimension. As the trend of three-dimensional models put into usage dominates in geophysics gradually, to gain a three-dimensional model, or even a four-dimensional model, with little hindrance is becoming more and more desperate and essential. In order to make this come true, a modelling system (the Interactive Gravity and Magnetic Application System, or IGMAS) is presented, which combines information from various geophysical methods by incorporating GIS functions. Thus, the underground area of our planet could be demonstrated much more intuitively, visually, and vividly.

In a nutshell, there is no denying that GIS has an enormous active influence on geophysics. Geophysical research requires the analysis of a lot of different types of data, such as satellite imagery, digital aerial photo mosaics, seismic surveys, surface geology studies, subsurface and cross section interpretations and images, well locations, and existing infrastructure information. A GIS can tie these data together to the location in question and allow you to overlay, view, and manipulate the data in the form of a map to thoroughly analyze the potential for finding new or extending play potential.

As discussed earlier, GIS technology assists geophysicists with the design of the sampling sites and in measuring lines outside in the field, better staying away from unreachable areas, and enables the geophysical data collection procedure to be easier and more efficient. GIS makes it more convenient to reserve and manage large geophysical projects with several datasets by building up a comprehensive database. It allows us to establish 2-D/3-D models of the subsurface area and works as a very powerful tool to synthesize different kinds of complex geophysical datasets to make their connections display intuitively for further analysis and interpretation (www.gisgeography.com). Whereas there are many limitations, including immature database construction, poor data quality, low data sharing degree, etc.

Regardless of the disadvantages, the application of GIS in geophysics will definitely have a wonderful, bright, shining future because opportunities and challenges always exist side by side.

14.3.2 APPLICATIONS OF GIS IN GEOPHYSICAL EXPLORATION SURVEY

14.3.2.1 Oil and Gas Exploration in Ethiopia Using GIS

In oil and gas exploration, geophysical applications like seismic methods, gravitational methods, and to a lesser degree, magnetic methods will determine and locate the geological structures suitable for the accumulation of oil. The applications of those methods will result into different approaches of the determination prospect area (Table 14.3).

The purpose of this study is to demonstrate the exploration workflow by utilizing GIS in designing a seismic survey with the integration of geological, geophysical, and other data to improve the exploration process in terms of cost, accuracy, and time. The Ethiopia study has demonstrated this by compiling, registering, and analyzing data obtained from various methods (remote sensing, geophysics, etc.) in GIS as part of the PCOSB Blocks in Ethiopia.

Through the integrated datasets using GIS, raster data, such as aerial photos or satellite imagery, can be incorporated with vector data, and surface culture, such as contours, topographic landmarks, digital elevation model, or points of interest, can be presented. Maps of the following features and characteristics were created in the first phase of exploration reconnaissance mapping program (Latef and Kahar, 2008):

- Topography maps inclusive of satellite images, digital elevation model (DEM) and triangular irregular network (TIN)
- Cultural maps of Ethiopia from public data and internal datasets
- Geophysical maps inclusive of gravity, magnetic, and geological datasets
- Seismic survey design maps inclusive of 2-D and 3-D

TABLE 14.3**Geophysical Methods**

Magnetic	Gravity	Seismic
Will determine:	Will determine:	Can tell:
1. Limits of basins	1. Basin shape and extent	1. Types of structures
2. Depth to basement	2. Structural trends	2. Structural character
3. Basement lineations	3. Fault locations	3. Relation of different features
	4. Prospects and leads	4. Sediment velocities
		5. Sediments which flow
		6. Unconformities
		7. Direction of sediment source
		8. Elements of geological history
		9. Inferences from reflection character
		10. Inferences of depositional environment
		11. Inferences of age
		12. Problems in mapping
		13. Prospect leads and prospects definition
		14. Leads as to gas accumulation

14.3.2.2 GIS as Project Manager in Geophysical Applications Software, Quaid-i-Azam University, Islamabad, Pakistan

A practical example is included to demonstrate the implementation of GIS component in a seismic refraction software which is a three-stage application for picking arrival times, computing refractor model, and finally, calculating statics (Khalid et al., 2011). A new project is created, and all project datasets and processing parameters are defined into its database. Using GIS or Project Explorer, any dataset can be interactively opened and processed according to the predefined job sequence (Khalid et al., 2011).

14.3.2.3 Contribution of GIS in Mineral Exploration through Mineralization Distribution Characterization: Application to the Sétifien/ Hodna Massifs and the Eastern Saharan Atlas (NE Algeria)

The integration and processing of geological/depositional, geophysical, and geochemical data in a geographic information system contributed to the accurate determination of the spatial distribution and particularities of the Pb-Zn-Fe-Ba (Cu, Hg, F, Sr) mineralization of the Sétifien/Hodna region and the eastern Saharan Atlas (Ysbaa et al., 2019). The concentrations of Pb-Zn-Cu-Fe (%) in the database were processed in ArcInfo and merged with the spatial database in the GIS. Interpolation maps show different colours that represent different variability intervals of Pb-Zn-Cu-Fe (%) (Ysbaa et al., 2019).

14.3.2.4 Preparation of Geospatial Data Infrastructure in Geological Survey of India: A Rewarding Experience with BRGM of France: Tons Valley Base Metal Prospect, Uttar Pradesh

A large amount of Earth science data collected by the Geological Survey of India since its inception in 1851 has been brought to electronic media for storage, quick access, retrieval, and analysis in GIS environment, through a collaboration with the BRGM of France, starting in 1996. In the initial stage, ten data domains were identified, out of which five, viz., map 50,000, geochemical exploration, mineral exploration, drilling, and coal, were developed and implemented for preparing geospatial database on Oracle RDBMS platform. Subsequently, data pertaining to the remaining domains,

viz., geophysics, photogeology and remote sensing, rock analysis, natural hazards, glaciology, and environment, were developed and implemented.

Multivariate analysis pertaining to Tons Valley base metal prospect was taken up. Multivariate parameters, viz., topographic contours, streams, chemical data (Cu, Pb, Zn, Ni, Ag, and Co), geophysical data (chargeability, self-potential, and resistivity), and photogeological data (lineaments and land-use/geology), were processed and used for multicriteria analysis. The SYNCOMPO software, and superimposed in GIS environment, was used to generate an output showing six parameters individually, and the result was draped over the digital elevation model (Srivastava et al., 1996). Two significant anomalies were obtained; one coincided with the known zone of mineralization, while the second anomaly was recommended for drilling, which indicated mineralized zone in depth. The study proved the efficacy of multivariate analysis for mineral targeting using data layers in GIS environment (Srivastava et al., 1996).

14.3.2.5 GSI Seismic/Earthquake Studies, Aizawl Town, Mizoram, in GIS Environment

Seismic and landslide hazard assessment of the Aizawl town, Mizoram, was taken up. A landslide hazard assessment map was prepared in this connection, which indicated that 5% of the area is under high hazard, 41.3% of the area is under moderate hazard, and 53.7% of the area is under low hazard categories, respectively. The spatial distribution of the respective hazard zones thus obtained will aid in land use planning strategies and disaster mitigation. An integrated seismic hazard assessment map using different parameters was generated on a GIS platform, which revealed that 13% of the area is under relatively very high hazard, 39.1% of the area is under relatively high hazard, 42.3% of the area is under relatively moderate hazard, and 5.6% is under relatively low hazard categories, respectively (GSI, 2016, p. 1,068).

REFERENCES

- Aitken, M., 1974. *Physics and Archaeology*. 2nd edn. Oxford: Clarendon Press.
- Allard, P., Behncke, B., D'Amico, S., Neri, M. and Gambino, S., 2006. Mount Etna 1993–2005: Anatomy of an evolving eruptive cycle. *Earth Science Reviews*, Elsevier, Netherlands, vol. 78, pp. 85–114.
- Alparone, S., Cocina, O., Gambino, A., Mostaccio, A., Spampinato, S., Tuve, T. and Ursino, A., 2012. Seismological features of the Pernicana-Provenzana fault system (Mt. Etna, Italy) and implications for the dynamics of northeastern flank of the volcano. *Journal of Volcanology and Geothermal Research*, Elsevier, Netherlands, vol. 251, pp. 16–26.
- Argus, D.F. and Heflin, M.B., 1995. Plate motion and crustal deformation estimated with geodetic data from the global positioning system. *Geophysical Research Letters*, vol. 22, pp. 1973–1976.
- Arnadottir, T., Beavan, J. and Pearson, C., 1995. Deformation associated with the 18 June, 1994, Arthur's Pass earthquake, New Zealand, NZ. *Journal of Geology and Geophysics*, vol. 38, pp. 553–558.
- Banerjee, P., Burgmann, R., Nagarajan, B. and Apel, E., 2008. Intraplate deformation of the Indian subcontinent. *Geophysical Research Letters*, vol. 35, p. L18301.
- Batayneh, A., 2011. Archaeogeophysics—archaeological prospection—a mini review. *Journal of King Saud University—Science*, vol. 23, no. 1, p. 83089.
- Beavan, J., 1993. Surface deformation due to the Mw 7.8 Guam earthquake of August 8, 1993. *Eos*, Transactions, American Geophysical Union, vol. 74, p. 183.
- Beazeley, G., 1919. Air photography in archaeology. *Geographical Journal*, vol. 53, pp. 330–335.
- Bennett, R.A., Reilinger, R.E., Rodi, W., Li, Y., Toksoz, M.N. and Hadnut, K., 1995. Coseismic fault slip associated with the 1992 Mw 6.1 Joshua Tree, California, earthquake: Implications for the Joshua Tree-Landers earthquake sequence. *Journal of Geophysical Research*, vol. 100, no. B4, pp. 6443–6461.
- Berlin, G., Ambler, J., Hevly, R. and Schaber, G., 1977. Identification of Sinagua agricultural fields by aerial thermography, soil chemistry, pollen/plant analysis and archaeology. *American Antiquity*, vol. 42, pp. 588–600.
- Binford, L., 1964. A consideration of archaeological research design. *American Antiquity*, vol. 29, pp. 425–441.
- Blewitt, G., Heflin, M.B., Hurst, K.J., Jefferson, D.C., Webb, F.H. and Zumberge, J.F., 1993. Absolute far-field displacements from the 28 June 1992 Landers earthquake sequence. *Nature*, vol. 361, no. 6410, pp. 340–342.

- Bock, Y., Agnew, D.C., Fang, P., Genrich, J.F., Hager, B.H., et al., 1993. Detection of crustal deformation from the Landers earthquake sequence using continuous geodetic measurements. *Nature*, vol. 361, no. 6410, pp. 337–340.
- Bonaccorso, A., Bonforte, A., Currenti, G., Del Negro, C., Di Stefano, A. and Greco, F., 2011. Magma storage, eruptive activity and flank instability: Inferences from ground deformation and gravity changes during the 1993–2000 recharging of Mt. Etna volcano. *Journal of Volcanology and Geothermal Research*, Elsevier, Netherlands, vol. 200, pp. 245–254.
- Bonaccorso, A., Bonforte, A., Guglielmino, F., Palano, M. and Puglisi, G., 2006. Composite ground deformation pattern forerunning the 2004–2005 Mount Etna eruption. *Journal of Volcanology and Geothermal Research*, Elsevier, Netherlands, vol. 111, p. B12207.
- Bonaccorso, A., Currenti, G. and Del Negro, C., 2013. Interaction of volcano-tectonic fault with magma storage, intrusion and flank instability: A thirty years study at Mt. Etna. volcano. *Journal of Geophysical Research*, vol. 251, pp. 127–136.
- Bonforte, A., Bonaccorso, A., Guglielmino, F., Palano, M. and Puglisi, G., 2008. Feeding system and magma storage beneath Mount Etna as revealed by recent inflation/deflation cycles. *Journal of Geophysical Research*, vol. 113, pp. B05406.
- Bonforte, A., Guglielmino, F., Coltelli, M., Ferretti, A. and Puglisi, G., 2011. Structural assessment of Mount Etna volcano from permanent Scatterers analysis. *Geochemistry, Geophysics, Geosystems*, American Geophysical Union (AGU), United States, vol. 12.
- Bonforte, A. and Puglisi, G., 2003. Magma uprising and flank dynamics on Mount Etna volcano, studied using GPS data (1994–1995). *Journal Geophysical Research*, vol. 108, no. B3.
- Borgia, A., 1994. Dynamic basis of volcanic spreading. *Geophysical Research Letters*, vol. 99, no. B9, pp. 17781–17804.
- Borgia, A., Ferrari, L. and Pasquare, G., 1992. Importance of gravitational spreading in the tectonic and volcanic evolution of Mount Etna. *Nature*, vol. 357, pp. 231–235.
- Borgia, A., Lanari, R., Sansosti, E., Tesauro, M., Berardino, P., Fornaro, G., Neri, M. and Murray, J.B., 2000. Actively growing anticlines beneath Catania from the distal motion of Mount Etna's decollement measured by SAR interferometry and GPS. *Geophysical Research Letters*, vol. 27, no. 20, pp. 3409–3412.
- Branca, S. and Ferrara, V., 2013. The morphostructural setting of Mount Etna sedimentary basement (Italy): Implications for the geometry and volume of the volcano and its flank instability. *Tectonophysics*, vol. 586, pp. 46–64.
- Chatteraj, S.L., Prasad, G., Sharma, R.U., Champati ray, P.K., van der Meer, F.D., Guha, A. and Pour, A.B., 2020. Integration of remote sensing, gravity and geochemical data for exploration of Cu-mineralization in Alwar basin, Rajasthan, India. *International Journal of Earth Observation and Geoinformation*, vol. 91, p. 102162.
- Cortes, A., Garduno, V., Macias, J., Navarro-Ochoa, C., Komorowski, J., Saucedo, R. and Gavilanes, J., 2010. Geologic mapping of the Colima volcanic complex (Mexico) and implications for hazard assessment. In: Groppelli, G., Viereck-Goette, L. (eds), *Stratigraphy and Geology of Volcanic Areas*. Geological Society of America special paper, vol. 464, pp. 251–266.
- DeMets, C., Gordon, R.G., Argus, D.F. and Stein, S., 1990. Current plate motions. *Geophysical Journal International*, vol. 101, pp. 425–478.
- DeMets, C., Gordon, R.G., Argus, D.F. and Stein, S., 1994. Effect of recent revisions to the geomagnetic reversal time scale on estimates of current plate motions. *Geophysical Journal Letters*, vol. 21, pp. 2191–2199.
- Drewett, P.L., 2003. *Field Archaeology: Introduction*. London: Routledge.
- Dvorak, J., 1994. An earthquake cycle along the south flank of Kilauea Volcano, Hawaii. *Journal of Geophysical Research*, vol. 99, no. B5, pp. 9533–9541.
- Ebert, J., 1984. Remote sensing applications in archaeology. *Advances in Archaeological Method and Theory*, vol. 7, pp. 293–362.
- Frey Mueller, J., King, N.E. and Segall, P., 1994. The co-seismic slip distribution of the Landers earth quake. *Bulletin of the Seismological Society of America*, vol. 84, no. 3, pp. 646–659.
- Gahalaut, V.K., Kundu Bhaskar, Singh Laishram Sunil, Catherine Joshi, Arun Kumar, Devchandra, M., Tiwari, R.P., Samanta, S.K., Ambikapathy, A.P., Mahesh, J.K., Bansal, A. and Narsaiah, M., 2013. Aseismic plate boundary in the Indo-Burmese wedge, northwest Sunda Arc. *Geology*, vol. 41, pp. 235–238.
- Goetz, A.F.H., Vane, G., Solomon, J.E. and Rock, B.N., 1985. Imaging spectrometry for earth remote sensing. *Science*, vol. 228, no. 4704, pp. 1147–1153, DOI: 10.1126/science.228.4704.1147.

- GSI, 2014. Remote sensing and aerial surveys. e-News, Special Issue, RSAS. Geological Survey of India, Headquarters, Kolkata.
- GSI, 2016. Institutional report. *Proceedings Indian National Science Academy*, vol. 82, no. 3, July Spl Issue 2016, pp. 1061–1081. DOI: 10.16943/ptinsa/2016/48504.
- Guglielmino, F., Bignami, C., Bonforte, A., Briole, P., Obrizzo, F., Puglisi, G., Stramondo, S. and Wegmuller, U., 2011. Analysis of satellite and in situ ground deformation data integrated by the SISTEM approach: The April 3, 2010 earthquake along the Pernicana fault (Mt. Etna—Italy) case study. *Earth Planet Science Letters*, vol. 312, pp. 327–336.
- Gupta, R.P., 2003. Introduction. In: *Remote Sensing Geology*. 2nd edn. New York: Springer- Verlag Berlin Heidelberg GmbH, pp. 1–16, DOI: 10.1007/978-3-662-05283-9.
- Howard, A. et al., 2015. Remote sensing and habitat mapping for bearded capuchin monkeys (*Sapajus libidinosus*): Landscapes for the use of stone tools. *Journal of Applied Remote Sensing*, vol. 9, no. 1, p. 096020. DOI: 10.1117/1.JRS.9.096020.
- Hudnut, K.W., Bock, Y., Cline, M., Fang, P., Feng, Y., et al., 1994. Co-seismic displacements of the 1992 Landers earthquake sequence. *Bulletin of the Seismological Society of America*, vol. 84, pp. 625–645.
- Hudnut, K.W., Shen, Z., Murray, M., McClusky, S. and King, R., 1996. Co-seismic displacements of the 1994 Northridge, California, earthquake. *Bulletin Seismological Society America*, vol. 86, no. 1B, pp. S19–S36.
- Jain, R., Sharma, R.U. and Kumar, A., 2017. Mineral mapping techniques: A case study in southeastern Rajasthan. Conference Paper: 38th Asian Conference on Remote Sensing, New Delhi, October.
- Khalid, A.K., Gulraiz, A. and Zulfiqar, A., 2011. GIS as project manager in geophysical applications software. *Journal of Geographic Information System*, vol. 3, pp. 166–172.
- Kruse, F.A., 2002. Mineral mapping with AVIRIS and EO-1 hyperion. *Proceedings of the 11th JPL Airborne Geoscience Workshop*. Pasadena, CA: Jet Propulsion Laboratory Publication.
- Kruse, F.A., Boardman, J.W. and Huntington, J.F., 2003. Comparison of airborne hyperspectral data and EO-1 hyperion for mineral mapping. *IEEE Transactions on Geoscience and Remote Sensing*, vol. 41, no. 6, pp. 1388–1400, DOI: 10.1109/TGRS.2003.812908.
- Lal, B.B., 1949. Sisupalgarh 1948: An early historical fort in eastern India. *Ancient India*, vol. 5, pp. 62–105.
- Larsen, S., Reilinger, R., Neugebauer, H. and Strange, W., 1992. Global Positioning System measurements of deformations associated with the 1987 Superstition Hills earthquake. *Journal of Geophysical Research*, vol. 97, no. 4, pp. 4885–4902.
- Larson, K.M. and Frey Mueller, J., 1995. Relative motions of the Australian, Pacific and Antarctic plates estimated by the global positioning system. *Geophysical Research Letters*, vol. 22, no. 1, pp. 37–40.
- Latef, M.Z.B.A. and Kahar, R.B., 2008. *Oil and Gas Exploration in Ethiopia using GIS*. Kuala Lumpur, Malaysia: GIS Development.
- Lingyan, R., Yanning, Z., Wei, W. and Qilin, Z., 2017. A hyperspectral image classification framework with spatial pixel pair features. *Sensors*, vol. 17, no. 10, p. 2421, DOI: 10.3390/s17102421. PMC 5677443. PMID 29065535.
- Lisowski, M., Prescott, W.H., Savage, J.C. and Johnston, M.J.S., 1990. Geodetic estimate of coseismic slip during the 1989 Loma Prieta, California, earthquake. *Geophysical Research Letters*, vol. 17, no. 9, pp. 1437–1440.
- Lundgren, P.R., Berardino, P., Coltelli, M., Fornaro, G., Lanari, R., Puglisi, G., Sansosti, E. and Tesauro, M., 2003. Coupled magma chamber inflation and sector collapse slip observed with synthetic aperture radar interferometry on Mt. Etna volcano. *Journal of Geophysical Research*, vol. 108, no. B5, p. 2247.
- Lundgren, P.R., Casu, F., Manzo, M., Pepe, A., Berardino, P., Sansosti, E. and Lanari, R., 2004. Gravity and magma induced spreading of Mount Etna volcano revealed by satellite radar interferometry. *Geophysical Research Letters*, vol. 30, p. 1388.
- Lundgren, P.R., Wolf, S.K., Protti, M. and Hurst, K.J., 1993. GPS measurements of crustal deformations associated with the 22 April 1991, Valle de la Estrella, Costa Rica earthquake. *Geophysical Research Letters*, vol. 20, no. 5, pp. 407–410.
- Lyons, T. and Avery, T., 1977. *Remote Sensing: A Handbook for Archaeologists and Cultural Resource Managers US Government Printing Office*. Washington, DC: Cultural Resources Management Division, National Park Service (NPS).
- Mahesh, J.K., Catherine Joshi, Gahalaut, V.K., Kundu, B., Ambikapathy, A.P., Bansal, A.L., Premkishore Narsaiah, M., Ghavri, S., Chadha, R.K., Choudhary, P., Singh, D.K., Singh, S. K., Kumar, S., Nagarajan, B., Bhatt, B.C., Tiwari, R.P., Kumar, A., Kumar A., Bhu, H. and Kalita, S., 2012. Rigid Indian plate: Constraints from the GPS measurements. *Gondwana Research*, vol. 22, pp. 1068–1072.

- Melbourne, T., Carmichael, I., DeMets, C., Hudnut, K., Sanchez, O., et al., 1996. Shallow faulting and regional subsidence in the Mw 8.0 1995 Jalisco Subduction earthquake. *Geophysical Research Letters* (Submitted).
- Mishra, D.R., Ogashawara, I. and Gitelson, A.A., 2017. *Bio-Optical Modeling and Remote Sensing of Inland Waters*. Waltham, MA: Elsevier Inc.
- Mitrovica, J.X., Davis, J.L. and Shapiro, II, 1994. A spectral formalism for computing three-dimensional deformations due to surface loads, 1: Theory. *Journal of Geophysical Research*, vol. 99, pp. 7057–7074.
- Miyazaki, S., Tsuji, H., Hatanaka, Y., Abe, Y., Yoshimura, A., et al., 1996. Establishment of the nationwide GPS array (GRAPES) and its initial results on the crustal deformation of Japan. *Bulletin of the Geographical Survey Institute (Japan)*, vol. 42, pp. 27–41.
- Murray, J.B., van Wyk de Vries, B., Sargent, P. and Wooller, L., 2018. Gravitational sliding of the Mt. Etna massif along a sloping basement. *Bulletin of Volcanology*, vol. 80, no. 40.
- Murray, M., Marshall, G.A., Lisowski, M. and Stein, R.S., 1996. The 1992 M D 7 Cape Mendocino, California, Earthquake: Coseismic deformation at the south end of the Cascadia mega thrust. *Journal of Geophysical Research*, vol. 101, pp. 17707–17725.
- Neri, M., Acocella, V. and Behncke, B., 2004. The role of the Pernicana fault system in the spreading of Mt. Etna (Italy) during the 2002–2003 eruption. *The Bulletin of Volcanology*, vol. 66, pp. 417–430.
- Neri, M. and Rossi, M., 2002. Geometria e volume dell'apparato vulcanico etneo: il contributo offerto dall'uso di mappe digitali. *Quaderni di Geofisica*, vol. 20, pp. 1–16.
- Norini, G. and Acocella, V., 2011. Analogue modeling of flank instability at Mount Etna: Understanding the driving factors. *Journal of Geophysical Research*, vol. 116, p. B07206.
- Nunnari, G. and Puglisi, G., 1994. Ground deformation studies during the 1991–1993 Etna eruption using GPS data. *Acta Vulcanologica*, Italy, vol. 4, pp. 101–107.
- Obrizzo, F., Pingue, F., Troise, C. and De Natale, G., 2001. Coseismic displacements and creeping along Pernicana ault (Mt. Etna) in the last seventeen years: A detailed study of a structure on a volcano. *Journal of Volcanology Geothermal Research*, vol. 9, pp. 109–131.
- Obrizzo, F., Pingue, F., Troise, C. and De Natale, G., 2004. Bayesian inversion of 1994–1998 vertical displacements at Mt Etna: Evidence for magma intrusion. *Geophysical Journal International*, vol. 157, pp. 935–946.
- Owen, S., Segall, P., Frey Mueller J., Miklius, A. and Denlinger, R., 1995. Rapid deformation of the south flank of Kilauea Volcano, Hawaii. *Science*, vol. 267, pp. 1328–1332.
- Perisset, M. and Tabbagh, A., 1981. Interpretation of thermal prospection on bare soils. *Archaeometry*, vol. 23, pp. 169–189.
- Pippi, I., 1989. Mineral identification by the AVIRIS data. Geoscience and Remote Sensing Symposium, IGARSS'89: 12th International Canadian Symposium on Remote Sensing, Vancouver, Canada, pp. 944–947.
- Prabhakar, V.N. and Korisettar, R., 2017. Ground survey to aerial survey: Methods and best practices in systematic archaeological explorations and excavations. *Current Science*, vol. 113, no. 10, pp. 1873–1890.
- Pu, R., 2017. *Hyperspectral Remote Sensing: Fundamentals and Practices (Remote Sensing Applications Series)*. Kindle edn. Taylor and Francis Series in Remote Sensing Applications, Qihao Weng, Series Editor, p. 490.
- Puglisi, G., Briole, P. and Bonforte, A., 2004. Twelve years of ground deformation studies on Mt. Etna volcano based on GPS surveys. In: Bonaccorso, A., Calvari, S., Coltelli, M., Del Negro, C., Falsaperla, S. (eds), *Mt. Etna: Volcano Laboratory*. Geophysical Monograph Series, 143. Washington, DC: AGU, pp. 321–341.
- Qingxi, T., Yongqi, X. and Lifu, Z., 2014. Progress in hyperspectral remote sensing science and technology in China over the past three decades. *IEEE Journal of Selected Topics in Applied Earth Observations and Remote Sensing*, vol. 7, no. 1, pp. 70–91.
- Ruegg, J.C., Campos, J., Armijo, R., Barrientos, S., Briole, P., et al., 1996. The Mw D 8:1 Antofagasta (North Chile) earthquake of July 30, 1995: First results from teleseismic and geodetic data. *Geophysical Research Letters*, vol. 23, pp. 917–920.
- Sabins, F.F., 1999. Remote sensing for mineral exploration. *Ore Geology Reviews*, vol. 14, no. 3–4, pp. 157–183, DOI: 10.1016/S0169-1368 (99)00007-4.
- Schowengerdt, R.A., 2007. *Remote Sensing: Models and Methods for Image Processing*. 3rd edn. San Diego, CA: Academic Press, p. 2, ISBN 9780123694072.

- Segall, P. and Davis, J.L., 1997. GPS applications for geodynamics and earthquake studies. *Annual Review of Earth and Planetary Sciences*, vol. 6, no. 25, pp. 301–36, DOI: 10.1146/annurev.earth.25.1.301.
- Shen, Z., Ge, B.X., Jackson, D.D., Potter, D., Cline, M. and Sung, L., 1996. North ridge earthquake ruptures based on the Global Positioning System measurements. *Bulletin of Seismological Society America*, vol. 86, no. 1B, pp. 37–48.
- Siebert, L., 1992. Threats from debris avalanches. *Nature*, vol. 356, pp. 658–659.
- Srivastava, G.S., Malaviya, A.K., Srivastava, R. and Kulshrestha, A.K., 1996. Preparation of geospatial data infrastructure in Geological Survey of India: A rewarding experience with BRGM of France. Technical Commission IV, WG—IV/1 (data source: Unpublished G.S.I. Report for F.S. 1996–1998).
- Stringer, W. and Cook, J., 1974. *Feasibility Study for Locating Archaeological Village Sites by Satellite Remote Sensing Techniques: Report to the US National Aeronautics and Space Administration for Contract No. NAS5–21833, Task 110-N*. Washington, DC: NASA Technical Reports Server (NTRS).
- Swayze, G., Clark, R.N., Kruse, F., Sutley, S. and Gallagher, A., 1992. Ground-truthing AVIRIS mineral mapping at Cuprite, Nevada. In: *JPL, Summaries of the Third Annual JPL Airborne Geoscience Workshop*, vol. 1, AVIRIS Workshop. JPL Publication, pp. 47–49.
- Tabei, T., Kato, T., Catane, J.P.L., Chachin, T., Fujimori, K., et al., 1996. Crustal deformation associated with the 1995 Hyogo-ken Nanbu earthquake, Japan, derived from GPS measurements. *Journal of Physics of the Earth*, Special issue, In press.
- Thakuria, T. et al., 2013. Google Earth as an archaeological tool in the developing world: An example from India. *Society for American Archaeology Archaeological Record*, pp. 20–24.
- Ting-ting, Z. and Fei, L., 2012. Application of hyperspectral remote sensing in mineral identification and mapping. 2nd International Conference on Computer Science and Network Technology, Changchun, China, pp. 103–106.
- Tiwari, V.M., Veenadhari, B., Gahalaut, V.K., Mukherjee, S. and Dimri, V.P., 2014. Gravity, GPS and geomagnetic data in India. *Proceedings of the Indian National Science Academy*, vol. 80, no. 3, pp. 705–712, DOI: 10.16943/ptinsa/2014/v80i3/55145.
- Tsuji, H., Hatanaka, Y., Sagiya, T. and Hashimoto, M., 1995. Co-seismic crustal deformation from the 1994 Hokkaido-Toho-Okai earthquake monitored by a nationwide continuous GPS array in Japan. *Geophysical Research Letters*, vol. 22, no. 13, pp. 1669–1672.
- Waddington, H., 1946. Adilabad: A part of the ‘fourth’ Delhi. *Ancient India*, vol. 1, pp. 60–76.
- Walter, T.R., Acocella, V., Neri, M. and Amelung, F., 2005. Feedback processes between magmatic events and flank movement at Mt. Etna (Italy) during the 2002–2003 eruption. *Journal of Geophysical Research*, vol. 110, p. B10205.
- Williams, C., Arnadottir, T. and Segall, P., 1993. Coseismic deformation and dislocation models of the 1989 Loma Prieta earthquake derived from Global Positioning System measurements. *Journal of Geophysical Research*, vol. 98, pp. 4567–4578.
- Wooller, L., van Wyk de Vries, B., Murray, J.B., Rymer, H. and Meyer, S., 2004. Volcano spreading controlled by dipping substrata. *Geology*, vol. 32, no. 7, pp. 573–576.
- Xiaoran, L., Amelung, F., Shao, Y., Ye, S., Liu, M. and Xie, C., 2020. Rheology of the zagros lithosphere from post-seismic deformation of the 2017 Mw7.3 Kermanshah, Iraq, Earthquake. *Remote Sensing*, vol. 12, no. 2032, pp. 1–14, DOI: 10.3390/rs12122032.
- Ysbaa, S., Rachid, N., Omar, H., Abdelhak, B. and Madjid, C., 2019. Contribution of GIS in Mineral Exploration through Mineralization Distribution Characterization: Application to the Sétifien/Hodna Massifs and the Eastern Saharan Atlas (NE Algeria). Crimson Publishers, Research Article, *Aspects in Mininig & Mineral Science*, vol. 3, no. 2, pp. 373–381. AMMS.000556.2019, DOI: 10.31031/AMMS.2019.03.000556.

15 Marine Geophysics

15.1 INTRODUCTION

Marine geophysics is a very important scientific discipline in the geophysical field of exploration in which a detailed study is being generally made for deciphering geology distinctly of the seafloor and sub-seafloor due to sufficient physical property contrasts, generating different geophysical signatures, recorded by different measuring devices of different geophysical methods during marine surveys of the oceans using gravity, magnetic, swath bathymetry, seismic reflection, and electrical resistivity to address local dispositions of anomalous bodies under the sea as well as to get deep insight into the regional and global issues, such as seafloor spreading, continental drift, and plate tectonics. It is reported that seafloor sediments keep the memory of the ocean circulation and climate change for a longer period of geological timescale. In order to interpret marine geophysical data into the corresponding geology, the principal branches of geophysical tools are required, as, for example, ocean floors, as well as the midoceanic ridge, are found to be the best source for heat flow data. Marine geophysical observations aim at understanding the nature and behaviour of the seabed and the underlying geological structure. Geophysical data are important not only from the scientific point of view but also for a number of practical applications that involve the research of minerals and energy and the industry more or less associated with natural resources.

The Geological Survey of India is one of the pioneer organizations involved in exploring the Indian Ocean using different geophysical sensors. The Marine and Coastal Survey Division (M&CSD) of GSI (erstwhile Marine Wing [MW]/Offshore Mineral Exploration and Marine Geology [OME&MG]) started coastal and marine surveys in a miniscule manner way back in the early 1970s by surveying and mapping of selected parts of coastal waters using country boats and other small fishing crafts. The work is mainly confined to the collection of surface sediment samples and spot bathymetric data.

The Marine and Coastal Survey Division (M&CSD) of the Geological Survey of India has carried out 738 cruises within the EEZ of India since 1983 using three research vessels—RV *Samudra Manthan*, RV *Samudra Shaudhikama*, and RV *Samudra Kaustubh*—to generate baseline marine geoscientific data. Marine geophysical surveys are deployed for seabed mapping to detect and delineate heavy minerals (HM) in the Territorial water (up to 12 nautical miles) in the Indian Ocean for demarcating the prospective economic resources for the future. Similarly, the estimation of construction-grade sand off Kerala has been delineated. Further, vast resources of high-grade calcium carbonate mud (lime mud) that occur along the outer continental shelf and continental slope off Gujarat and Maharashtra are being evaluated in detail.

With the acquisition of one ocean-going vessel, RV *Samudra Manthan*, in 1983, and two coastal launches, RV *Samudra Kaustubh* and RV *Samudra Shaudhikama*, in 1984, and with RV *Samudra Ratnakar* (www.gsi.gov.in), since 2014, having been conceived as a replacement vessel for the ageing RV *Samudra Manthan* (RVSM), GSI commenced a systematic survey of the seabed within the exclusive economic zone (EEZ) and in international waters. In pursuance of the Charter of the Geological Survey of India, the main emphasis of the marine survey has been on seabed mapping and exploration for nonliving resources in the offshore sector in the EEZ and in international waters. On the basis of the delineated offshore, obvious geological potential (OGP) and focused offshore mineral exploration (FOME) cruises are being taken up for polymetallic deposits (PMD) that include rare Earth element (REE), coplatinum group of elements/metals (PGE/PGM), gold, etc., phosphorites and lime mud, heavy mineral placers, and construction-grade sand. During 2012–2015, the Marine and Coastal Survey Division estimated total resources of 132,196 mt of placer deposits over an area of 466 km² along the east and west coasts of India (GSI, 2016).

The various thrust areas are identified to cater to the needs and challenges of the nation and fulfil the short-term and long-term socio-economic needs. They encompass baseline data collection (geological and geophysical), assessment of offshore mineral potential, geochemical scan for hydrocarbons, basic research and development in furtherance of understanding of various processes of Earth science, industrial and port development, evaluation of natural processes, and thereby, mitigating the damages due to natural hazards, historical, and archaeological evolution of coastal areas.

15.2 PROPAGATION OF SEISMIC WAVES

The controlled source of marine survey involves the detonation of a sound source within the water column, wherein acoustic energy gets generated in the form of sound waves, which travel spherically in a similar way when a stone is tossed into a river or pond, generating wavelets-like whirling of waves outwardly from the point where the stone was tossed. The fundamental principles involved were first formulated by Christian Huygens in the 17th century. The seismic source and characteristics of the media of propagation (i.e. water type and depth) determine the basic character of the seismic waves generated. These waves have amplitude (strength), frequency, and velocity (speed).

The relation between frequency and wavelength is given by

$$\lambda = \frac{v}{f}$$

where λ is the wavelength (metres), v is the velocity (metres per second), and f is the frequency (hertz) of a propagating acoustic wave. It is the wavelength that determines the resolution of a particular seismic system. Two types of seismic waves, namely, compressional or primary P waves and shear or S waves, are used in geophysical surveys of an area. The P waves move away from the source and reflectors by alternately compressing and dilating the elastic media (water or sediments) along the travel path. The primary wave (V_p) is related to the elastic moduli (namely, bulk modulus and rigidity modulus) as follows:

$$V_p = \sqrt{\frac{k + \frac{4}{3}\mu}{\rho}}$$

where V_p is the compressional wave velocity, k is the bulk modulus, μ is the rigidity modulus or shear modulus, and ρ is the density of the media through which the wave is traveling. The rigidity modulus and bulk modulus are proportionality constants relating to shear and stress strain for ideal homogenous and continuous elastic solids. The primary wave is a very significant source of seismic wave that finds applications in high-resolution geoengineering (HRG) reflection profiling techniques. The shear or S waves oscillate in a direction normal to (at 90°) the travel axis and are described mathematically by

$$V_s = \sqrt{\frac{\mu}{\rho}}$$

where V_s is the velocity of shear waves, ρ is the density, and μ is the rigidity modulus of the media through which they are traveling. As described in the beginning, the secondary or shear wave velocities are slower than those of the P waves because of the zero-rigidity modulus, in which the particle motion is perpendicular to the direction of propagation; consequently, the S wave does not propagate through the water column or media consisting of water. Such characteristics of S wave make it unfit to be used for high-resolution geoengineering (HRG) profiling methods.

The mathematics involved into three-dimensional elastic wave propagation is not so easy; therefore, the theory of wave path is applied to understand the processes of 3-D elastic wave propagation. This theory is described by Huygens's principle, whereby every point on a wave front is the source of a new wave that also travels out from it in a spherical manner. If the spherical waves have a large radius, they can be treated as planes, and lines perpendicular to the wave fronts, called wave paths or rays, can be used to describe the propagation of seismic waves.

The intensity and decay of a transmitted acoustic signal (sound wave) are commonly described using the decibel scale, where the change in intensity of the sound wave due to purely spreading loss from the source is given by the equation

$$I_{dB} = 10 \log_{10} \frac{I_2}{I_1}$$

where I_{dB} is the intensity of the signal in decibels and I_1 and I_2 are the linear intensity measured at two locations. The preceding equation is used to describe signal strength such that a -10 dB drop in the intensity means a drop in the intensity by a factor of 10, given the logarithmic nature of the decibel scale. Therefore, a -20 dB reduction is a drop in the intensity by a factor of 100, and a -30 dB drop is a drop by a factor of 1,000. An important parameter that helps characterize the seismic source and its impact on marine mammals is the sound pressure level (SPL). The SPL, with units of decibels (dB), in water is defined as

$$SPL = 20 \log_{10} \frac{p}{p_{ref}}$$

As per the guidelines of the National Marine Fisheries Service (NMFS) and the Marine Mammal Protection Act:

Level A harassment: sound levels >180 dB re $1 \mu\text{Pa}$ (RMS)

Level B harassment: sound levels >160 dB re $1 \mu\text{Pa}$ (RMS)

15.3 SOURCES OF SEDIMENTS

In order to understand the geotechnical properties of marine sediments, the study of soil mechanics is of paramount importance since the static and dynamic forces are dictated by the physical properties of marine sediments and the corresponding response of a sedimentary system. As per the study made by Moore and Keller (1984), the following are the types of marine sediments:

Terrigenous. It is derived from lands. Dominated by volume, covering 45% of the seafloor.

Thus, terrigenous sediments in the deep ocean entail turbidites and debris near margins.

However, the turbidites can travel about 1,000 km from the continental shelf.

Biogenic. Derived from shells/skeletons of organisms—siliceous, calcareous (second most important by volume, covering 55% of the seafloor). Consists of two main oozes in particular. Siliceous oozes consist of diatoms and radiolarians. *Diatoms* are single-celled algae found at high latitudes. *Radiolarians* are amoeba-like and are found in equatorial regions.

Calcareous oozes consist of coccoliths, foraminifera, and pteropods. Foraminifera are amoeba-like, and pteropods are drifting molluscs that are most easily dissolved.

Hydrogenous. Manganese and phosphorite nodules, hydrothermal sediments, evaporites (salt).

Cosmogenic. Derived from meteorites, tektites, and cosmic spherules.

Based on laboratory or in situ test results, the civil engineer attempts to predict the behaviour of soils and the structure they support. For example, he or she may determine the amount of settlement

that will take place following the emplacement of a structure on the seafloor or assess the necessary depth to which piles should be driven to support a platform subjected to a designed maximum load (e.g. drilling rig superstructures, processing facilities, storm wave forces, winds, etc.).

Ideally, the selection of an appropriate borehole site for obtaining core samples for geotechnical analyses should be predicated upon the local geology, which is determined from the interpretation of an HRG survey. The detailed review and correlation between the survey data and geotechnical test results provide the necessary information for the establishment of a sound foundation design within the offshore environment. The selection of borehole location is very crucial, and utmost care must be taken based on the interest of the HRG data rather than the frequent drilling of boreholes for exploratory study only.

15.4 GEOPHYSICAL TECHNIQUES AND INSTRUMENTS TO SURVEY THE OCEANS

A comprehensive marine geoengineering survey demands the deployment of a number of acoustic tools from a precisely positioned survey vessel. Geophysical equipment deployed includes echo sounders, side-scanning sonars, subbottom profilers, sparkers, or other deeper-penetrating acoustic sources. Total field intensity or gradient magnetometers and appropriate positioning equipment are also kept for use during marine surveys. The interpretation of such survey data permits the assessment of geological and other features both on and beneath the seafloor, some of which may constitute either hazards (such features may include faults; shallow accumulations of overpressured gas; the presence of man-made debris, such as lost drilling equipment or sunken vessels; or lateral variations of the physical properties of the seafloor, such as that occurring in association with infilled relict river channels) or constraints to the placement of bottom-supported or penetrating engineering structures. Lack of attention to such features may lead to the loss of a drilling platform by a blowout and the escape of large volumes of hydrocarbons into the marine environment. The high-resolution seismic profile of the figure (Trabant, 1984) illustrates the result of a blowout where a drilling platform was lost.

15.4.1 SIDE-SCAN SONAR SURVEY

A side-scan sonar system is an acoustic device that aims to produce a 2-D image of the seabed of near-photographic quality. In early surveys, side-scan sonar systems were used since the 1970s, but the systems were initially very expensive for a variety of purposes, including archaeological exploration. At that time, their resolution was only just high enough to detect and locate larger sites. Since then, technological advances have made side-scan sonars powerful enough and portable, requiring little power and being able to display detailed information about a wide variety of sites. Until the recent emergence of high-resolution multibeam sonars, it was the most commonly used geophysical system for maritime archaeology.

For conventional side-scan sonar, two sets of transducers (one port and one starboard) are carried by a streamlined tow fish, towed behind the survey vessel. The tow fish ensures that the transducers are far enough from the noise generated by the vessel, increases stability by reducing pitch and roll, and enables the system to 'fly' within a few metres of the seabed. Each transducer transmits a fan-shaped acoustic pulse (ping) perpendicular to its travel path. The transducers act as both source and receiver, which means that the transducer alternately switches between emitting and registering signals. The frequencies emitted can range from 100 Hz (long-range, tens of kilometres) to over 1 MHz (short-range, in metres). More commonly, side-scan systems used for surveying in shallow water produce a frequency between 50 kHz (medium-range, hundreds of metres) and 500 kHz (short-range, tens of metres), with pulse lengths of tens to hundreds of seconds. Systems that operate at two frequencies simultaneously (e.g. 100 kHz and 500 kHz) are being increasingly used. The generated side-scan beams are narrow in the horizontal plane (along-track direction) and wide in the vertical plane (across-track direction) (Fish and Carr, 1990). For the dual-frequency system,

for example, the horizontal beam is smaller than 1° , and the vertical beam can vary between 40° and 60° . Such angles enable the sonar to ensonify a narrow strip of the seabed while the sonar transmits vertically into the entire water column.

As the beam propagates through the water, it will eventually interact with the seafloor or with objects on it. Most of the energy is reflected away from the side-scan sonar system as the result of specular reflection. A small portion of the energy is lost in the subsurface, and another small portion, known as backscatter, reflects back to the side-scan system. The amplitude of this returned signal is measured by the transducers, together with the travel time, amplified, recorded, and displayed as a time series.

Necessarily, data from different pings are stitched together to display a long continuous image of the seabed. The amount of backscatter is determined by three factors: the local morphology of the surface ensonified, the small-scale (submetre) roughness of the surface, and the material properties of the seafloor. The image is created by a black-to-white display of the strength of the returning energy. Traditionally, the stronger the returning signals, the darker the tonality, with a lack of returning energy (e.g. a shadow behind an object) being displayed as lighter tones. However, this was a product of the thermal printers commonly used with the side-scan systems of the time. Modern computer-based acquisition packages have no such limitation, so the current convention is to display shadows as black tones and strongly reflecting targets as light tones. The role of different parameters is found different for different purposes as, for example, the travel times give information about the distance travelled from the transducer to the seabed, which is referred to as the slant range, completely different from the horizontal distance between the sonar and the target. This configuration provides the archaeologist with imagery of semiquantitative information on the morphology of the seabed equivalent to aerial photography and also a first-order-magnitude description of the material variability of the seabed (i.e. the distribution of different sediment types). The figure (Figure 6, page 16 in *Marine Geophysics Data Acquisition, Processing and Interpretation Guidance Notes*, 2013; Fish and Carr, 1990) shows an example, with commonly used key terms.

15.4.2 SWATH BATHYMETRY SURVEY

Techniques to measure the ocean's depths changed significantly during the 20th century. It is demonstrated that the swath bathymetry is an important marine survey arrangement in which depths are measured in a line extending outwardly with respect to the sonar transducer. Systems obtain the data in the swath at right angles to the direction of the motion of the transducer head. As the head continues to move forward, the profile sweeps out in a ribbon-shaped surface of depth measurement thus known as the swath. Swath bathymetry systems have been tested from the 1960s and were first used commercially in the 1970s. However, only after the 1990s did the system gain popularity in the hydrographical and archaeological worlds. This section explains the principles of the acoustic bathymetric survey systems but mainly concentrates on swath techniques, as it is believed that these are now regarded as the standard survey tool for coastal and shelf surveying. There are currently three swath bathymetry systems that involve two different technologies to achieve bathymetry measurements across the seafloor.

15.4.2.1 Single-Beam Echo Sounders

These instruments measure the depth of the seabed by recording the time it takes for a sound pulse to travel from the transducer to the seabed and back (Fish & Carr, 1990). Based on simple physics, the mean speed of sound through water having velocity variation from 1,480 m/s to 1,500 m/s enables the estimation of depth using the two-way travel time (TWTT) from the relation of depth, which is equal to half of the TWTT time multiplied by the velocity of sound through water. The frequencies used by commercially available systems range from 10 kHz to 200 kHz. For high-resolution mapping, high-frequency systems with narrow beams (typically between 2° and 5°) need to be used in order for the circular footprint to cover a small enough area to obtain sufficient accuracy. The transducer, which acts both

as a source and a receiver, is normally mounted on either the bottom or the side of the vessel's hull and only measures the depth vertically beneath it. The narrow beam echo sounders might need beam stabilization (i.e. compensate for the roll, pitch, and heave of the vessel) in order to measure the depth vertically below the transducer. A single-beam echo sounder survey produces an image of the seabed topography along the track of the vessel. However, to produce a bathymetric map, several parallel profiles are needed, and thus, significant interpolation between profiles is always necessary.

The principle behind the swath bathymetry systems is to increase the seabed coverage and reduce survey time by using several beams at different angles to the vertical. Such systems are available in a range of frequencies, varying from 12 kHz to 500 kHz. The highest-frequency systems provide centimetre-scale images of the seabed. Swath systems transmit a fan of ultrasonic sound, broad in the across-track direction (typically 120° to 150°) and narrow in the along-track direction (between 0.5° and 3°). The position of each echo can be computed from the angle and the travel time (i.e. the range) of the returned signal. Depending on how the angle and travel time pairs are determined, two different systems are recognized: beam-forming multibeam echo sounders and interferometric or phase discrimination sonars.

15.4.2.2 Multibeam Sonar

This system employs a technique known as beam-forming to ascertain the depth of the seafloor (Fish & Carr, 1990). The system is made up of two transducer arrays, a transmitting array whose long axis is parallel to the direction of travel, and a receiving array perpendicular to that. Each array produces a fan-shaped beam that is narrow in the direction of its long axis. The arrays are made up of a number of identical and equally spaced transducer elements, forming a fixed number (e.g. 126, 254, or 512) of transmitted and received beams at different angles (hence the name 'multibeam'). Through a process called 'beam steering', the receiving array can be altered so that echoes from a number of directions can be received. Using beam steering, each receiver beam intersects the emission beam, resulting in a series of 'footprints' on the seabed along the ensonified area. The echo arrival time of each footprint and the angle of the received beam, corresponding to that footprint, is then used to determine the depth of the seabed. Using this system, the seabed is sampled more densely at small angles than at higher angles. Therefore, the accuracy and resolution will be highest for the inner parts of the swath and will decrease with increasing swath width (Fish and Carr, 1990). The beam-forming process cannot distinguish between multiple travel times corresponding to a single angle. However, these multiple reflectors are often much weaker than the primary reflectors, hence in practice cause few problems.

15.4.2.3 Interferometric Sonar

This setup features two sonar heads mounted on a V-shaped framework (Fish & Carr, 1990). Each sonar involves one transmitting array and at least two receiving arrays that would be set up parallel to each other in the direction of travel. It may be noted that similar to side-scan sonar, the transmitting transducer arrays produce a single beam that is wide in the vertical direction and narrow in the horizontal direction. The receiving arrays, spaced at carefully chosen fixed distances, detect the backscattered signal at different arrival times. The travel times provide the range to the echo, while the phase difference measured between the signals at the different receivers determines the angle of arrival. The knowledge of both the range and the angle makes it possible to calculate the exact position of the echo. However, the interferometric system cannot distinguish between multiple angles with the same travel times. Consequently, in theory, the system will struggle to produce an image of a steep seafloor or of complicated (upstanding) structures correctly. Manufacturers are developing ways to deal with this shortcoming, for example, by assessing the strength of the reflected data.

In contrast to the multibeam system, interferometric sonar receives thousands of beams. The density of the sounding locations is larger for the outer parts of the swath than for the inner parts (Fish and Carr, 1990). The high density of data points is reduced during the postprocessing by

calculating a mean grid. The interferometric sonar is also frequently referred to as a bathymetric side-scan sonar, for the system can accurately record the amplitude (or backscatter) of the returned signal, and the data can be treated as a side-scan record as well as offering bathymetric information. Mitchel and Smith (2014) indicate the comparison of bathymetric data density.

15.4.3 SUBBOTTOM PROFILER SURVEY

It is a proven fact that the high-resolution seismic reflection systems operating at acoustic frequencies between 1.0 and 14.0 kHz are grouped within the general category of subbottom profilers. These systems allow for continuous high-resolution seismic profile recordings of the uppermost 30 m of strata, below the seafloor. The subbottom profiler (SBP) has long been a principal tool in the high-resolution investigation of offshore sites for the emplacement of pipelines and bottom-supported structures, such as jack-up rigs; in the reconnaissance of sites for the construction of artificial islands for nuclear power plants and petroleum drilling operations; and in the exploration for shallow mineral deposits, such as sands and gravels, as well as numerous other applications.

In the case of marine surveys, the instrumentation is generally a towed transmitter and an array of geophones (Figure 15.1). Setting up of the field equipment on the ocean bottom or floor with specialized array geometry is required for data acquisition. The greatest challenge in the marine data acquisition is associated with the survey profiles/lines that have to be kept in straight line as much as possible from the research vessel that remains on a body of unstable water, which demands complex equipment for deep imaging of structures beneath the ocean floors. The research vessel is equipped with GPS sensors for getting information on the locational positions of the equipment that tows with the research vessel. The marine seismic method maintains a constant offset equal to the distance from the GPS

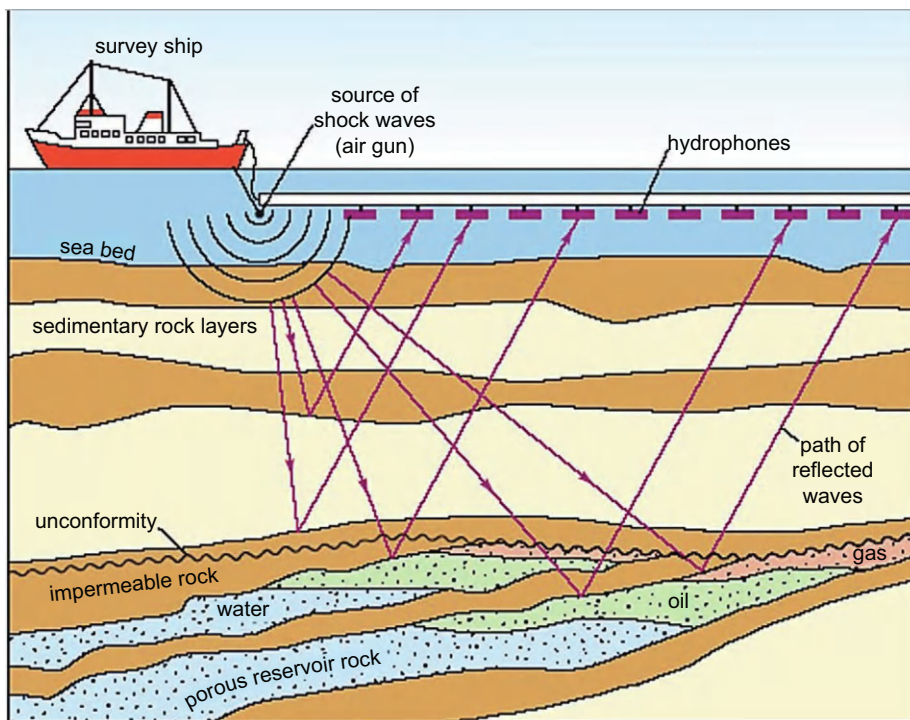


FIGURE 15.1 The swath bathymetry model adapted from an educational website on petroleum illustrates the basic application of seismic data acquisition in a marine setting.

Source: Drake (1970).

sensor to the geophone array. As discussed earlier, the controlled source marine seismic surveys may use a variety of seismic sources available in the market, such as water guns (20–1,500 Hz), airgun (100–1,500 Hz), sparkers (50–4,000 Hz), boomers (300–3,000 Hz), and chirp systems (500 Hz–12 kHz, 2–7 kHz, 4–24 kHz, 3.5 kHz, and 200 kHz). The greatest resolution of near-surface structure is generally obtained from the higher-frequency sources, such as the chirp systems, while the lower frequency tends to better characterize structure at depth (USGS case study in the San Francisco Bay).

The technique of the subbottom profiler is similar to that used by single-beam echo sounders, but at lower frequencies, the sound waves penetrate the seafloor, whilst at boundaries between layers of different acoustic impedance (i.e. the product of sound velocity and density), changes in seismic energy occur, where a part of the seismic energy reflects back to a detector, part of the energy is transmitted through the boundary to deeper layers, and part of the energy is lost through scattering. It is the impedance contrast between the layers that determines the amount of energy that gets reflected back. The transceiver or a separate acoustic receiver (i.e. a hydrophone) is towed directly behind the seismic source, which detects the reflected energy. The subsurface image is the result of the amplitudes (two-way travel times) of the reflected waves. All surface-towed subbottom data must be tidally corrected, and unless real-time kinematic (RTK) data have been acquired, this can be done using tide gauge information. When a grid of data has been acquired, the crossing points between different lines will indicate whether the tidal correction has been applied correctly. In studies, the seismic reflection profilers have shown that the active continental margins usually show undisturbed sediments in the bottoms of the trenches. The seismic reflection profiling plays a huge role in locating potential oil and natural gas deposits.

15.4.4 MULTICHANNEL AND 3-D SEISMIC REFLECTION SURVEYING

The subbottom profilers discussed earlier (Section 15.4.3) assume the simplest and most common method of marine seismic reflection surveying: single-channel profiling in which a marine seismic/acoustic source is towed behind the vessel, triggered at a fixed rate, and the signals reflected from the sedimentary column detected by a hydrophone streamer made of several receiver elements towed close to the source. The signals detected by the individual receiver elements are summed, and a single trace is recorded.

Advances in subbottom techniques, mainly triggered by the offshore industry, have led to the development of multichannel and 3-D systems. These systems are now being adapted and are slowly finding their way into shallow-water surveying. During multichannel reflection seismic surveying, a long hydrophone streamer made up of individual receivers is towed behind the source. Instead of instantly summing the received signals, each individual seismic signal received at each section is recorded as a separate trace. This way, recordings of reflected pulses are obtained at several offset distances from a shot point. This data can then be used to estimate the velocity of sound through the sedimentary column, which, in turn, can be used to convert the two-way travel time to true depths. After processing and stacking data points with a common position, the resulting profile will have an improved signal-to-noise ratio, a higher horizontal trace density, and an improved ability to discriminate among reflection events in comparison with traditional single-channel data. The figure [Figure 2.3, Page 2.33 in Bureau of Ocean Energy Management (BOEM) Executive Summary, Fugro; Volume-2, Geophysical Surveys Benefits and Risks] shows the schematic of water bottom multiples in seismic reflection data and 3-D configuration of offshore seismic data acquisition system, respectively.

15.4.5 GRAVITY/MAGNETIC SURVEY

The gravity process is the force of gravitational attraction, g , which is exerted by the Earth on an object. The surface is not constant, but it all varies from place to place. It is well documented that the variation in gravity value is controlled by the shape of the Earth (e.g. oblate spheroid). The purpose of gravity measurements is to assess the zone associated either with mass deficiency or mass sufficiency

and then translate the gravity signature into reasonable geological structures. There can be many different interpretations of gravity measurements that can conclude important features in the data. Large negative anomalies associated with deep-sea trenches must be associated with corresponding deficiencies of mass. With the advent of geophysical instrumentation and advancement in processing analytics, a well-defined marine gravity measurement upgraded to use vessels mounted with different kinds of equipment, such as gravimeter, magnetometers, since early years till 1960. Great improvements in measuring accuracy are achieved by the development of control electronics and, above all, enhanced navigation data gained by global navigational satellite systems, like GPS. With the advent of new technology, now we are able to calculate gravity from satellite altimetry measurements, resulting in global datasets from which free air gravity anomaly values for nearly all oceanic areas of the Earth have now been assimilated, which, however, also demonstrates the constraints in the gravity data derived from satellite measurements relating to their limited spatial resolution and the addition of the systematic errors, especially in shallow water areas. In addition, for detailed modelling, good water depth data are needed and, whenever available, seismic data. It is therefore realized that in addition to these data, a better quality of the gravity data can be acquired at sea only by deploying research vessels.

In the past, marine gravity measurements seem to be a critical problem at first instance with the aim of measuring gravity changes with an accuracy of 1 mGal. As on vessels already, during moderate sea state conditions, the disturbing inertial accelerations reach up to 0.1 Gal, the gravity signal has to be separated from a noise signal, which is about 100,000 times higher. The Federal Institute for Geosciences and Natural Resources (BGR), Hannover, Germany, is running for this purpose the marine gravimeter system KSS32M of Bodensee Gravimeter Geosystem GmbH (BGGs). The KSS32M consists basically of the gravity sensor GSS30, which was installed on a gyro-stabilized platform (BGR—Marine gravimeter system KSS32M with the gravity sensor on the gyro-stabilized platform), and it has been reported that entire electronic units, including the power supply, remain integrated in the platform. The measured values, however, are not useable without further corrections of data through rigorous processing, such as:

1. In order to compensate for time delay because of inevitable damping, a time shift analysis to the measured data is to be carried out.
2. The value of the measured gravity changes because the speed around the Earth's rotation axis varies according to the east–west velocity component of the ship; gravity is a total sum vector of both the attraction of the Earth and the centrifugal force of its rotation. This effect is called the *Eötvös effect*. To measure gravity with an accuracy of 1 mGal, it is necessary to know the ship's speed in the east–west direction with accuracy better than 0.1 knots, which is now possible because of the Global Positioning System (GPS) that has come to be used, as earlier it was very difficult to achieve such high degree of accuracy in gravity measurement using shipborne surveys. The following formula shows the Eötvös correction in the unit of mGal:

$$\Delta g = 2\omega v \sin\theta \cos\varphi \times 0.514 \times 10^5 \approx 7.49 v \sin\theta \cos\varphi$$

where ω is the angular velocity of the rotation of the Earth ($= 7.29 \times 10^{-5} \text{ rad s}^{-1}$), v is the ship's velocity in knot (1 knot = 0.514 m s^{-1}), θ is the ship's direction measured from the north, and φ is the latitude of the observation station.

3. Subtraction of the normal gravity value, whereby the latitude-dependent gravity of a distinct reference ellipsoid (here GRS 80) as reference datum for the gravity anomalies is defined.

It has already been discussed that Mother Earth has appreciable magnetic bearing metals in its different subsurface strata, and the geomagnetic field persists because of the continuous

rotation of the viscous materials in the outer core based on the concept dynamo in physics. It is therefore realized that measurements at sea of the Earth's magnetic field must provide significant information about the interior of the Earth. In advanced marine magnetic surveys, measure the magnitude of the total intensity of the magnetic field. In contrast to the acoustic methods, magnetometers do not transmit any signals but rather measure geographical variations in the geomagnetic field. The total intensity of the magnetic field at the Earth's surface varies from 24,000 nT in equatorial regions to 66,000 nT at the poles; the geomagnetic field is $\sim 50,000$ nT overall in the UK. Variations within the magnetic field are caused by a number of factors: solar activity may cause variations of ~ 20 nT; geological features of a few nT to several hundreds of nT; nongeological ferromagnetic metallic objects, as for example, cannons or any buried metallic wreckages from ship or crashed aeroplanes trapped within the seabed can contribute additionally to the measured magnetic values up to few tens of nT; and it has also been reported that if metallic hulls are present beneath the seabed, then the magnetic values may surge by adding further up to thousands of nT. The measured intensity of metallic artefacts on or buried within sediments depends on the material, size, shape, depth of burial, and distance to the magnetometer. As the strength of the magnetic field is inversely proportional to the cube of the distance from the source, the magnetometer needs to pass the object as closely as possible and needs to be sufficiently sensitive to detect smaller, archaeologically significant objects. The main survey field parameters that must be determined before carrying out a magnetometer survey are (1) the type of magnetometer utilized, (2) the number and geometry (i.e. linear arrays, horizontal and/or vertical gradiometers), (3) the spacing between adjacent magnetometer track lines, and (4) the maximum altitude of the sensor. Currently, three types of magnetometers are used for maritime surveys: proton precession magnetometers, overhauser magnetometers, and optically pumped magnetometers.

Proton precession magnetometers were almost always used, until recently. The sensor component of this system is a cylindrical container filled with some liquid rich in hydrogen atoms. The container is surrounded by a coil connected to a power supply, amplifier, and frequency counter. When no current is running through the coil, the hydrogen protons align parallel to the ambient geomagnetic field. As a DC (continuous) current runs through the coil, a magnetic field is produced that is larger than and in a different direction from the geomagnetic field. This induced magnetic field causes the hydrogen protons to align along the direction of the applied field. The current is then turned off, and the protons return to their original alignment by spiralling, or 'processing', around the Earth's total magnetic field. This precession produces a time-varying magnetic field that, in turn, produces a small alternating current in the coil. The frequency of this AC (alternating) equals the frequency of the precession of the nuclei, which is proportional to the strength of the total field. Hence, by measuring this frequency, the total magnetic field strength can be determined. Proton precession magnetometers are inexpensive but have a relatively slow sampling rate (0.5–2.0 s), with sensitivities of 0.2 nT to 1.0 nT. It is important that the sensor's orientation is at an angle with the Earth's magnetic field during surveying. Unfortunately, proton precession magnetometers are sensitive to heading errors, where the resulting measured total magnetic field over an object varies depending on the orientation of the tow fish.

Overhauser magnetometers are an improved proton precession magnetometer that has recently begun to replace traditional instruments. A special liquid containing free, unpaired electrons is combined with the hydrogen liquid in the sensor cell. The instrumentation of the overhauser magnetometers indicates that the sensor gets irradiated with a radio frequency magnetic field; consequently, unbound electrons transfer their energy to the hydrogen protons, which is why the resultant precession signals have a higher signal-to-noise ratio than for a normal proton precession magnetometer. The overhauser magnetometer sensitivity is $0.015 \text{ nT}/\sqrt{\text{Hz}}$, with an absolute accuracy of 0.1–0.2 nT. Furthermore, the signals are nondecaying, which means that the polarization and signal measurement can occur simultaneously, leading to an increased sampling rate (between 1 and 5 readings per second). It has also been demonstrated that the lightness, easy

portability, compactness, less power consumption, less prone to error in measurement make the overhauser instrument preferable. It is, however, the job of users and their purposes to decide what kind and type of magnetometers or any instruments they want to deploy for their geophysical surveys.

Optically pumped magnetometer is the third type of magnetometer and has an even higher precision, with a sensitivity of $0.004 \text{ nT}/\sqrt{\text{Hz}}$ and an absolute accuracy $<2 \text{ nT}$. Its depolarization is extremely rapid, up to 40 samples per second. The operational process is known as optical pumping. An optically pumped magnetometer comprises a glass cell containing a vapour of alkali atoms (rubidium, caesium, or potassium) and a polarized light source of the same element. Normally, electrons of these alkali atoms are positioned at two energy levels. As the polarized light is emitted through the vapour, electrons are bumped from these two levels to a third level. However, electrons at level 3 are not stable and will spontaneously decay back to levels 1 and 2. When level 1 is fully populated (and level 2 is depleted), the cell becomes transparent and the absorption of the polarized light stops. At this point, radio frequency (RF) power is applied to move the electrons back from level 1 to level 2, making the cell opaque again. The frequency of the RF field required to populate level 2 is a function of the ambient magnetic field. By measuring the light modulation as an effect of polarization and depolarization, together with the RF, the total magnetic field can be determined. It has been shown that the most commonly used alkali atom for marine survey is of caesium, which is why the nomenclature of magnetometer is coined as the caesium vapour magnetometer. Magnetometers are mostly used in combination with side-scan sonar. On land, diurnal, solar-caused variations of the Earth's magnetic field to the data can easily be corrected by placing a base station magnetometer at a fixed location in the vicinity of the survey area and collecting the magnetic variations continuously. The recorded survey data can then be adjusted using the base station's observations.

Contrary to land-based magnetic survey, the shipborne magnetic survey in the sea or ocean is very complex in the sense of generating accurate data because of chances of erroneous data correction in the sea and ocean due to an unstable platform, complicated survey logistics, and very high costs. Instead, diurnal corrections can be assessed by planning the survey such that certain lines cross each other. The diurnal variation can be calculated by studying the variation in magnetic field intensity at the crossover points. This method can only be used if the data points are collected at exactly the same depths above the seabed.

By using magnetic gradiometers, regional and temporal variations are automatically removed from the data. A *gradiometer* measures the gradient, or first spatial derivative, of the magnetic field. Two sensors, separated by a fixed distance, both measure the total field strength. The difference in field intensity measured by the two sensors, divided by their separation distance, results in a linear estimate of the gradient of the ambient field. The sensors can either be separated vertically or horizontally. The gradient resolves complex anomalies of shallow magnetic features into their individual elements, which can give information about the location, shape, and depth of an object. Despite the advantages and the availability of gradiometers, they are currently only rarely used by the marine surveying community. A more detailed review is provided by Camidge et al. (2010).

15.4.6 MARINE RESISTIVITY SURVEY

In the case of marine surveys, the instrumentation is generally a towed current electrode, or transmitter, and an array of potential electrodes or receivers (Figure 15.2). Field setup in the marine resistivity is very complicated because of the unsurmountable difficulty associated with selecting a straight profile for marine-based electrical surveys while in a vessel; besides, electrical imaging of the deep structures requires robust and heavy field equipment that poses another problem of logistics on the research or exploratory vessel. The figure (environmental geophysics—marine resistivity methods) demonstrates the deployment of a marine resistivity acquisition system. The additional components necessary for data processing are GPS, depth finders, and

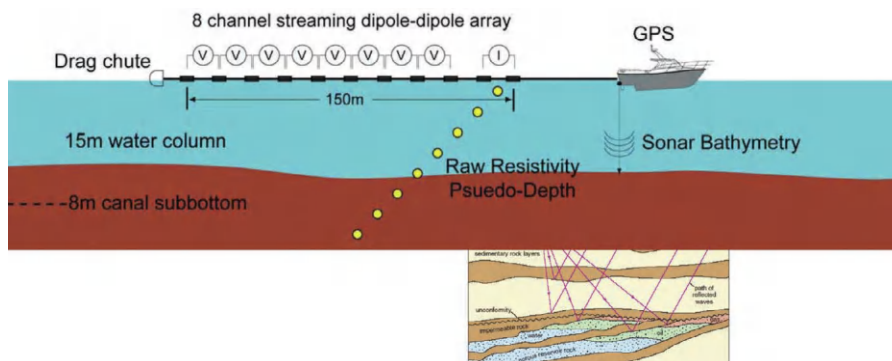


FIGURE 15.2 An illustration adapted from an educational website on petroleum illustrates the basic application of seismic data acquisition in a marine setting.

Source: Environmental geophysics—marine resistivity methods.

temperature recording devices. Many manufacturers offer specialized acquisition and processing software that integrates all the aforementioned data types. The Global Positioning System (GPS) provides positional location of the observation sites/stations, which works on mounting on the exploratory vessel towing the array by maintaining a constant offset equal to the distance from the GPS sensor to the electrode array. The figure shows the resistivity data acquisition under the marine geophysical method.

15.5 CORRECTIONS ADOPTED FOR MARINE GEOPHYSICAL DATASETS

Marine geophysical dataset is subjected to an important correction for the moving platform, referred to as the Eötvös correction, as described in details under the gravity chapter (Chapter 3), that constrains the variation in the gravity due to the movement of the ship. Other tools, such as bathymetric charts, ground-truthing satellite, and altimetry-derived gravity with the aim of understanding seafloor magnetization for plate kinematic studies based on grid compilations, are used for detailed investigations of the seafloor, crust, and upper mantle. These data are especially essential in remote localities far from ports and coastlines, where researchers are less likely to revisit (Wessel and Chandler, 2011). Distributions of gravity, bathymetry, and magnetic tracks are shown in the figure in Chandler and Wessel (2012).

Chandler and Wessel (2012) have initiated an errata-based along-track error correction system for marine geophysical track-line data, and in order to assess its performance, they computed cross-over errors before and after applying the along-track corrections. A comparison of global crossover standard deviations for bathymetry, magnetic, and gravity anomalies shows that the along-track corrections produce substantial error reductions. Based on the marine geophysical surveys analyzed, they find a strong tendency among source institutions to omit total field gravity measurements and to submit datasets without first removing all navigation and geophysical outliers. They find that ~89% of magnetic anomaly datasets require recalculation using the latest reference fields, while the remaining magnetic tracks lack total field data necessary for anomaly recalculation. The tendency of source institutions to submit preliminary and, on occasion, unprocessed or erroneous data places a burden on the scientific community which results in data processing redundancy among data users. The 5,230 errata tables generated in this research serve to mitigate this redundancy and improve the utility of the world's largest collection of marine geophysical tracking data. The improvements in data quality accomplished via errata tables will allow previously unused cruise data to be reconsidered and, in general, to strengthen scientific conclusions based on NGDC tracking data.

15.6 SURVEY APPROACHES AND SCHEDULES

15.6.1 EXPLORATORY OR RESEARCH VESSELS

The selection of survey vessel is based on a variety of factors, including suitability for the type of survey, its size, and the duration of survey, whether 12-hour (daylight hours) or 24-hour operation; distance from shore and water depth; ability to control appropriate survey speeds and maintain course; and vessel stability, deck space, and overall condition. Survey vessels come in various sizes and should be selected based on suitability for the survey and working in anticipated sea states and weather conditions. It is observed that data quality is very much important and utmost care has to be taken to keep data safe; prior to operation of the vessel, precautionary measures for safe data acquisition are to be taken. It is so because the state of the sea, the weather conditions in the sea, and the size and stability of the vessel have tremendous impacts on the quality of the data. One could continue collecting certain types of data, as per example, using multibeam if necessary. Shallow towed hydrophone streamers are typically the most sensitive survey equipment to sea states.

15.6.2 LINE SPACING

Line spacing will affect the resolution and detail of the ground model, the ability to identify or define geologic features, and the interconnectivity of various features (e.g. buried palaeodrainage networks). Projects in areas with complex geology may derive more benefit from closer-spaced regional lines than areas with more uniform geologic conditions.

15.6.3 INTEGRATION OF TEMPORALLY DIFFERENT DATA

If surveying is performed using a phased approach, then consideration for temporally sensitive data should be taken into account for the potential effect of merging data collected at different times. Dynamic processes can modify the seafloor, and the magnitude of those effects may vary with respect to water depth, seafloor materials, and hydrodynamic conditions. The three most significantly affected types of data include bathymetric, side-scan sonar, and magnetometer data.

Changes in the seafloor due to erosion, deposition, or movement of bedforms (e.g. sand waves) will affect the bathymetry. This could result in apparent seam artefacts where swaths from surveys conducted at different times meet each other. Migration of bedforms will result in an offset of the seam featured at the swath boundaries. However, if movement direction and rate of features can be interpreted by two different bathymetric datasets, then this can provide valuable information used in evaluating potential sediment transport hazards (e.g. sand wave mobility hazards). If bathymetry data from two different surveys are combined, then edge or seam effects should be identified and described in order to avoid them from being mistaken for real features. It has been reported that the side-scan sonar data get affected in a similar manner as bathymetric data. Magnetometer data also have the potential to be affected if collected at different times and/or using different sensors. Gridding and contouring of magnetometer data collected using different sensors with different sensitivities and/or at different times will have seam effects if not addressed appropriately. Gridding and contouring of the residual magnetic field (deviation from the total and local field) should mitigate seam effects and allow anomalies to be observed in merged datasets.

15.6.4 SURVEY DATA PROCESSING BOTTLENECKS

During large survey programs, it becomes necessary to conduct data processing while the survey is ongoing, to monitor and assess the quality of the data, identify data gaps, and identify data that need to be recollected. In some cases, data interpretation may also be conducted during survey operations to support the planning of other activities. For example, the survey vessel may also be used to

conduct sediment grab sampling or biological surveys. Processing and interpretation of side-scan sonar and/or multibeam data may be necessary in order to select sampling/camera locations.

15.7 ROLE OF MARINE GEOPHYSICS IN THE EXPLORATION OF DEEP-SEA NODULES AND IN THE BLUE ECONOMY

The global population, which is expected to exceed 8 billion by 2030, will require 30% more water, 40% more energy, and 50% more food. More than 40% of the world's population lives within 100 km of the coastline, and it continues to be on the uptrend. The Organisation for Economic Cooperation and Development (OECD) estimates the gross value added to the global output from the oceans shall be ~US\$3.5 trillion in 2030 and shall support more than 40 million jobs. Considering the strategic importance of the oceans, the United Nations (UN) has allocated a total global exclusive economic zone (EEZ) of ~137 million km², including ~12.0, 11.3, 8.5, and 2.3 million km² to France, the United States, Australia, and India, respectively. Based on the foundations of the 2012 Rio +20 UN Conference on sustainable development, the global community announced its commitment to the Sustainable Development Goals (SDG) 2030, in which goal 14 relates to the sustainable development of the ocean resources, by which economic activity is in balance with the long-term capacity of the ocean ecosystems, remain resilient and healthy. The importance of sustained utilization of the ocean resources is stressed by the UN declaring 2021–2030 to be the Decade of Ocean Science for Sustainable Development (Nayak, 2023, 2024).

In recent years, the importance of marine geophysical surveys has been considered at policy and planning level of those countries that depend on traditional ocean and marine economy. Therefore, they took significant steps to transform the traditional ocean and marine economy to a sustainable blue economy by commercializing the deep-ocean missions of the country concerned, including India (Nayak, 2024), for which a couple of sound pillars are recognized and can be adopted: appropriate governance in the sustained utilization of ocean resources, planning for coastal and marine economies, strategic vision of deep-ocean mapping, technology of the integrated marine exploration and management beneath the ocean floor under the deep-ocean program, and monitoring and time-bound regulatory reforms. It has been realized and accepted that the blue economy requires resilient systems, which may involve persistent innovation and advancement in achieving integrated ecological, economic, and social prosperity with the economic development of the country. Hence, the fast-emerging blue economy paradigm with its components shown in Table 15.1 requires proper estimation of the size of the opportunity and the nature of the risks involved. But at the same time, several strategic planning techniques with respect to asset investment, investment framework, and huge capital investment are required for a sustainable ocean as an asset, as well as for the growth of blue industries. Most recently, the Ministry of Earth Sciences, Government of India (www.moes.gov.in/schemes/dom), geared up with six thematic areas of deep-ocean mission that comprise the development of technologies for deep-sea mining, manned submersible and underwater robotics, the development of ocean climate change advisory services, technological innovations for the exploration and conservation of deep-sea biodiversity, deep-ocean survey and exploration, energy and fresh water from the ocean that align with the blue economic policy for sustainable economic growth in line with several different national and global initiatives in accordance with their respective road maps to generate feasible outcomes as per the 14th goal of the UN's 17 Sustainable Development Goals, harnessing the potential of the blue revolution or the blue economy (Nayak, 2023). Special budgetary provisions have been made in which integrated geophysical exploration can be made for deep-ocean survey and exploration (Rajeevan, 2018). Subsequent to the leadership of the small island developing states (SIDS) and the coastal nations which advocate for *blue economy*, major countries, including Australia, Brazil, UK, United the States, Russia, and Norway, have developed their national ocean policy and established hierarchal institutions at federal and state levels for ensuring progress in the blue growth. India has also its national ocean policy, referred to as

TABLE 15.1**Components of the Ocean and Marine Economy to a Blue or Sustainable Economy**

Type of Activity	Ocean Service	Industry
Extraction of nonliving resources	Minerals	Seabed mining
	Energy	Oil and natural gas
		Renewables
Harvesting of living resources	Fresh water	Desalination
	Seafood	Fisheries
		Aquaculture
Ocean commerce	Marine biotechnology	Pharmaceuticals, chemicals
	Transport and trade	Shipping
		Port infrastructure and services
	Tourism and recreation	Tourism
		Coastal development

Source: Atmanand et al. (2019).

the SAGAR Policy. With the blue economies of the United States, China, and the European Union estimated to be ~US\$1.5, 0.1, and 0.5 trillion, respectively, it is expected that the implementation of several sets of endeavour under the blue economy could significantly contribute to India's vision of becoming a US\$10 trillion economy in the future.

The global map in Hein et al. (2013) shows the location of the polymetallic nodules permissive regions and the exclusive economic zones (EEZs). There are four potential identifiable best-known polymetallic nodules distributed in the Clarion–Clipperton zone (CCZ), Peru Basin, Penrhyn Basin (including the Cook Islands EEZ), and Central Indian Ocean Basin (CIOB). Geologically and geophysically, it has been established that the polymetallic nodules also remain jacketed between seamounts and ridges in much of the equatorial West Pacific and parts of other island and coastal nations with their EEZs that allow exploration within permissive nodule areas. Mining operations for polymetallic nodules in the deep ocean consist of the following major components: collector robot on the ocean floor, vertical riser system, surface mining vessel, bulk carrier vessels at sea, and processing plant on land (Figure 15.3). The role of geophysical exploration is of paramount importance, where we use multibeam and hydroacoustic mapping tools with several geophysical sensors, gravity, magnetic, and seismic sensor systems mounted on the hull of a vessel that provide information about ocean floor topography and water depth (bathymetry) and the characteristics of structural heterogeneity associated with hard and soft rocks, sediments, and nodules on the ocean floor or beneath the ocean floor. Deep-towed systems and autonomous underwater vehicles (AUVs) can get closer to the ocean floor and, thus, provide a higher resolution (smaller footprint), via remotely operated underwater vehicle (ROV).

15.7.1 OPPORTUNITIES FOR INDIA

India has ample opportunity for conducting marine geoscientific surveys as it has a long coastline of 7,517 kms with 1,208 islands and has at least 13 major ports, 200 notified minor and intermediate ports. The EEZ includes a 1.64 million km² area near the Indian mainland and Lakshadweep, and 0.66 million km² in the Andaman and Nicobar Islands area. As per documented facts, India is a key member in the Indian Ocean Rim Association (IORA), an intergovernmental organization comprising of 21 member states and 9 dialogue nations having the mandate of strengthening regional cooperation and sustainable development within the Indian Ocean region that opened avenues to G20

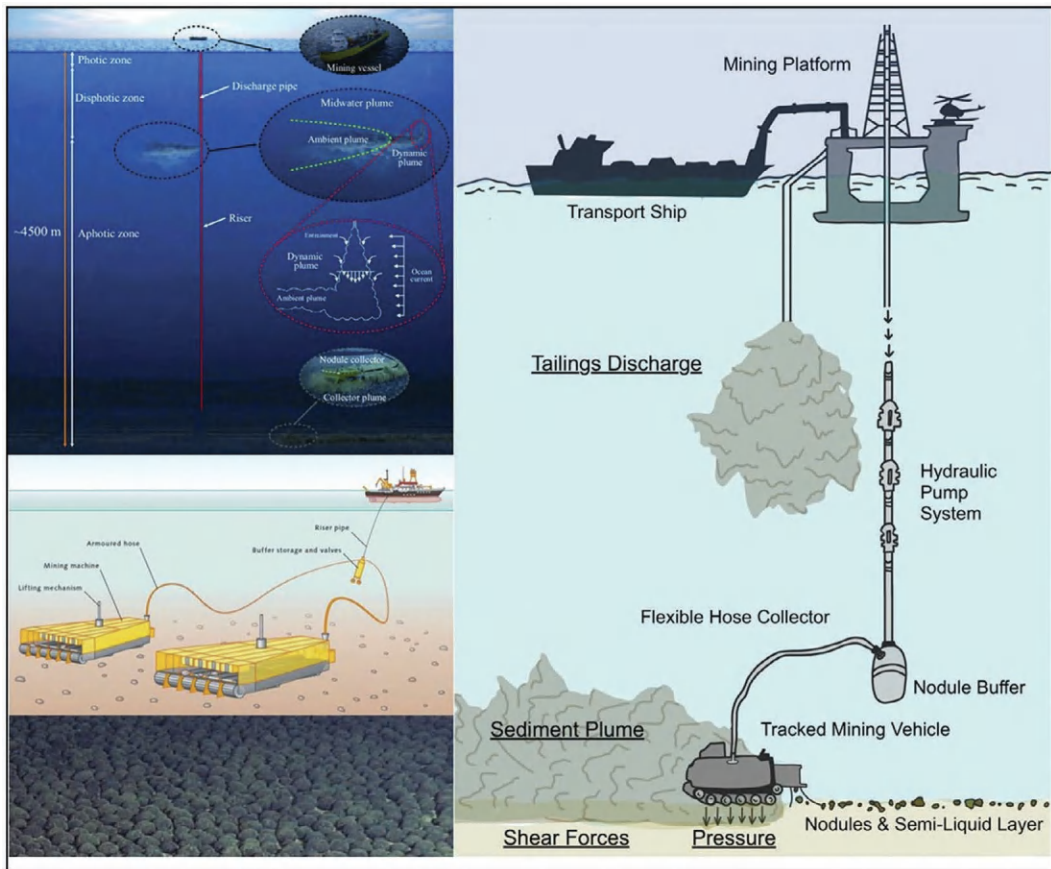


FIGURE 15.3 The components of a polymetallic nodule mining project.

Source: Kuhn et al. (2017); Sharma and Smith (2019).

countries towards sustainable consumption of the ocean, sea, and marine resources as one of the UN's Sustainable Development Goals, which can be achieved through extensive deep-ocean marine geophysical survey and exploration of mineral resources (Nayak, 2023). The IORA blue economy dialogue held in Goa in August 2015 passed the Goa declaration stressing the need to identify the thrust areas of the blue economy, which is now in the stage of implementation. During 2018, the National Institution of Transforming India (NITI), the Indian government's apex policy think tank, conducted discussions with all the stakeholders to identify the potential areas of the blue economy for blue growth of the country. Several initiatives have been initiated by the Ministry of Environment and Forests and Climate Change (MoE&FCC), the Ministry of Earth Sciences (MoES), the Ministry of Shipping, and other organizations. Some of the important initiatives include the deep-ocean mission, integrated coastal zone management, Sagarmala, Sagar, and Mausam, which emphasized on deep-ocean resource exploration, marine spatial planning, port development, maritime security, and improving the livelihood opportunities for coastal communities.

In order to synergize the multimistry efforts engaged in silos, seven working groups have been formed by the Government of India (Table 15.2) to formulate robust recommendations to capture the huge potential and opportunities in this sector. Thus, a sustainable, integrated, inclusive, and people-centric policy for blue economy is being evolved by India, and the policy statement states that '[the] blue economy refers to exploring and optimizing the potential of the oceans and seas which are under India's legal jurisdiction for socioeconomic development while preserving the health of the oceans'.

TABLE 15.2
Seven Working Groups under the Indian Government

National accounting framework for blue economy and ocean governance	Fisheries, aquaculture, and fish processing	
Coastal, deep-sea mining, and offshore energy	Manufacturing, emerging industries, trade, technology, services, and skill development	
Logistics, infrastructure, and shipping	Coastal marine spatial planning, tourism, and wellness	Security, strategic dimensions, and international engagements

Source: Atmanand et al. (2019).

15.8 DETAILED APPLICATIONS OF MARINE GEOPHYSICAL SURVEYS AND CASE HISTORIES

15.8.1 MARINE GEOLOGY AND COASTAL GEOPHYSICAL SURVEYS FOR SEABED MAPPING AND MINERAL RESOURCE EVALUATION IN INDIA

Detailed offshore surveys by GSI with R.V. Samudra Ratnakar have been carried out at high-resolution mapping (10 km interval grid pattern) of the EEZ of India with seismic, magnetic, gravity, and deep piston coring (Wadhawan, 2016). Obvious geologic potential areas in the offshore (OGP-offshore) have been identified and delineated based on reconnaissance surveys (Figure 15.4). Further detailed surveys in the OGP-offshore sectors have also been investigated with subbottom profiler, multibeam echo sounder (MBES), synthetic aperture mapping sonar, remotely operated vehicle, heat flow measuring system for detecting and delineating the mineral resource evaluation in ocean areas.

The Geological Survey of India conducted marine surveys in the Andaman–Nicobar region to assess the morphological setup of the Andaman–Nicobar trench and accretionary prism sediments using multibeam bathymetric surveys (MBBS) (Devas et al., 2015), and they showed that stabilization of the accretionary prism slices is supported by the near absence of the submarine slumps, whilst slices occur toward the east away from the trench axis. A successful shallow seismic investigation off North Tamil Nadu Coast, Bay of Bengal, in the Indian region has been carried out by Rao et al. (2016), and they inferred seabed and sub-seabed palaeomorphological features associated with palaeochannel on the inner shelf running parallel to the coast. A classical integrated analysis of magnetic, gravity, and multichannel seismic reflection data along a transect south-east of Sri Lanka, Bay of Bengal, has been carried out to unravel the structure and evolutionary history of the underlying sedimentary load and basement by Desa and Ramana (2021). Central Indian Ocean has been investigated extensively using gravity and magnetic methods to estimate the crustal depths (9–12 km) and the nature and extent of crustal deformation and corroborated with the estimates of Curie depth isotherm (Kumar and Begum, 2022). Sengupta et al. (2005) have already highlighted the temporal changes in magnetic anomaly around the Barren and Narcondam Islands, and they advocated that temporal changes in magnetic total field are responsible for the Curie isotherm depth (CID). Detection and forecast of Barren volcanic lava movement in Andaman–Nicobar Islands in Indian Ocean has been made by Satyanarayana et al. (2015) using marine-borne magnetic surveys, and they observed through repeat magnetic surveys that the Curie isotherm depth becomes maximum when the Barren volcano remains in the dormant state or inactive, whilst the continuous accumulation of magma in the magma chamber is manifested by the lessening or shallowing of Curie isotherm depth, suggesting the state of the next volcanic eruption or explosion that may occur in the future shortly. In the Bay of Bengal in the Indian Ocean (Radhakrishna et al., 2012) analyzed data acquired using integrated method consisted of shallow seismic reflection, magnetic,

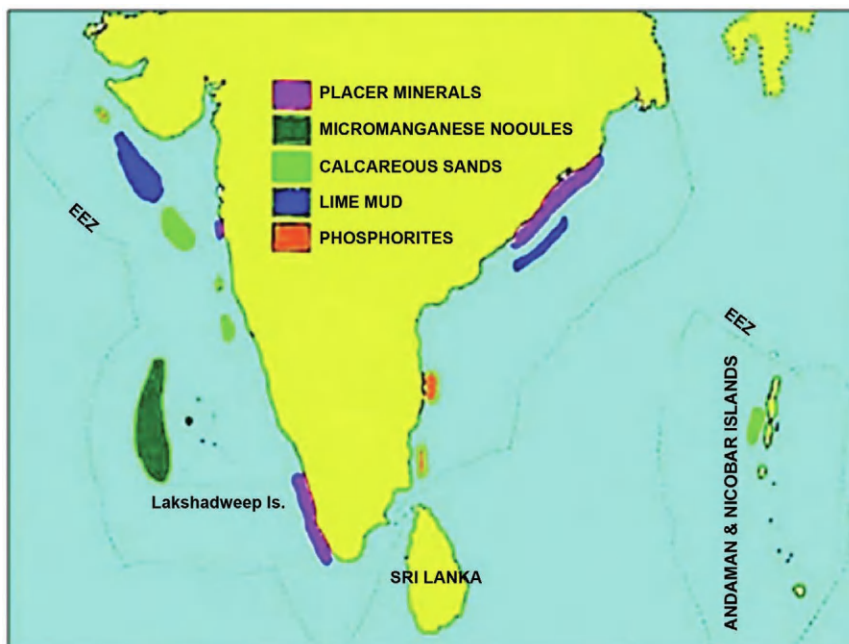


FIGURE 15.4 Offshore mineral-rich tracts constitute obvious geological potential areas for detailed exploration and scientific exploitation.

Source: Wadhawan (2016); GSI.

gravity methods. The data of free air satellite gravity and bathymetric surveys of the continental shelf, about 100 km off Sagar Island, have also been used to find out the seafloor configuration and variation of sediment thickness of the Late Quaternary period.

15.8.2 BASELINE GEOPHYSICAL DATA FOR HAZARD MANAGEMENT IN COASTAL AREAS IN RELATION TO EARTHQUAKES AND TSUNAMIS

A systematic study of geophysical data of the Eastern Continental Margin of India (ECMI) was taken up to identify the land–ocean tectonic (LOT) lineaments over the east coast of India and the possible neotectonic activity associated with the prominent structures, which have significant importance that help in delineating the offshore extension of some of the coastal lineaments (Murthy, 2016).

An integrated analysis of magnetic, gravity, and shallow seismic data coupled with reported earthquake data showed the occurrences of moderate seismicity are principally associated with some of these land–ocean tectonics (LOTs) of the Eastern Continental Margin of India (ECMI) (Murthy et al., 2010, 2012). It has also been reported that the coastal/offshore regions of Vizianagaram located north of Visakhapatnam and Ongole of the Andhra Pradesh margin and the Pondicherry shelf of the Tamil Nadu margin are associated with zones of weaknesses, where neotectonic activity has been established (Murthy et al., 2010, 2012; Murthy, 2016). The coastal and offshore areas of Visakhapatnam witnessed moderate seismicity of a magnitude lesser than and equal to 4.5, whilst the Puducherry offshore has witnessed occurrences of earthquakes of magnitude 5.5 (Murthy et al., 2002; Subrahmanyam et al., 2007). These results are useful in estimating the coastal stability of different parts of the east coast of India. Bathymetry data over the ECMI revealed the morphology of the shelf and slope of this margin, which in turn can be used as the baseline data for tsunami surge models. A comprehensive bathymetry contour and section map was generated for the entire ECMI (Murthy, 2016; Murthy et al., 2012), with details of shelf break, shelf width, and slope gradient at

different locations. Morphologically, the ECMI represents a mosaic of basinal and nonbasinal segments. The depth contours of the eastern continental margin nearly follow a north–south (N–S) trend in the southern part between Karaikal (10°55'N) and Nellore (14°30'N), from where they take a north–east–south–west (NE–SW) trend up to Paradip (20°15'N) in the north, thus running subparallel to the coastline (Murthy, 2016; Murthy et al., 2012). The contours indicate, in general, a narrow shelf in the south from Karaikal (10°55'N) to Nellore (14°30'N), also off the Krishna and Godavari river basins, and a relatively wider shelf to the north from Kakinada to Paradip. A detailed bathymetry map and sections of the Nagapattinam–Cuddalore shelf (from 10.5° to about 12°N) indicate (Murthy, 2011; Murthy et al., 2006) that one of the main reasons for the higher run-up heights and inundation in the Nagapattinam–Cuddalore coast (affected area marked in hatches in the figure in Murthy, 2016; Murthy et al., 2012, 2006) during the Indian Ocean tsunami of 26 December 2004 could be the concave shape of the shelf with a gentle slope, which might have accelerated the tsunami surge to flush through at a rapid force (Murthy, 2016; Murthy et al., 2006).

15.8.3 THE OUTER THAMES RIVER (ENGLAND) ESTUARY REGIONAL ENVIRONMENTAL CHARACTERISATION

The Outer Thames Estuary Regional Environmental Characterisation (REC) provides an environmental reference statement defining marine and seabed conditions within the study area (Ruth et al., 2013). A typical boomer seismic section reveals a buried paleochannel system in the Thames Estuary, which is situated up to 12 meters below the current seabed, based on data collected during the Outer Thames Estuary Regional Environmental Characterisation project.

15.8.4 DETECTION OF THE TORRE DEL GRECO SUBMERGED VOLCANIC STRUCTURE AT NAPLES BAY, SOUTHERN TYRRHENIAN SEA, ITALY

The three-dimensional reconstruction of the Torre del Greco structure has been carried out using seismic constraints and correlation of volcanic structure to bathymetry (Aiello et al., 2010). A kriging-type interpolation has been applied both to magnetic and to bathymetric data. The bathymetric dataset has been processed through the application of a median filter in order to remove the main sea-bottom corrugations by smoothing the sea-bottom topographic surface (Aiello et al., 2010). In this reconstruction, the sea-bottom topography offshore the Somma–Vesuvius volcanic complex is compared to the top of the Torre del Greco volcanic structure, represented by contour depth of corresponding seismic horizons. In conclusion, there is a good correspondence between the sea-bottom topographic surface and the top of the volcanic structure, which does not crop out at the sea-bottom.

15.8.5 GLACIAL AND GLACIALLY RELATED FEATURES ON THE CONTINENTAL MARGIN OF NORTH-WEST IRELAND

Four vast gully and canyon systems have been identified along the continental margin within the study area (Benetti et al., 2010). Intercanyon areas are characterized by three large sediment depositional lobes, frequent escarpments (especially around the gully and canyon heads), and shallow gullies (up to 40–50 m deep and 4 km wide), which are better developed below 1,500 m water depth. The northernmost depositional lobe is formed by glaciogenic debris flows deposited in the southern portion of the Donegal Fan (Stoker, 1995). Additional depositional lobes, offlapping onto the basin of Rockall Trough, are observed extending from the base of the continental slope to water depths of 2,400 to 2,600 m.

It is likely that the dimensions and location of many of these deep-water features are related to the presence of former ice margins on the outermost shelf or at the edge of the shelf that is prone to the delivery of large volumes of sediment and meltwater to the upper slope. In some areas, these sediments have been redistributed downslope by turbidity currents and mass wasting, thus contributing to further incision of the large canyon systems (Cronin et al., 2005; O'Reilly et al., 2007).

15.8.6 GRAVITY SURVEY AT THE OATES COAST OF ANTARCTICA DURING THE GERMAN ANTARCTIC NORTH VICTORIA LAND EXPEDITION (GANOVEX) VIII

A regional marine gravity survey was conducted during the 1999/2000 GANOVEX VIII/ItaliAntartide XV Expedition to the Oates Coast area, Antarctica. Precise values for the coordinates of each of the gravity stations are determined by differential GPS measurements (BGR). In this study, the measurements were tied to stations with known absolute gravity values, which were determined during previous GANOVEX expeditions to the area.

Free air anomaly map of the Oates Coast area is presented in the figure (gravity survey at the Oates coast during GANOVEX VIII, data source BGR). All existing gravity data from the expeditions GANOVEX IV to VIII (~310 points) have been used to produce this map. Their locations are plotted as small symbols. Strong negative gravity anomalies are dominantly visible over glaciers. The reason for the negative gravity anomalies is the enormous mass deficits due to the much lower densities of the ice (which is not corrected for in free air), and probably underlying sediments, compared to the density of the bedrock. A remarkable, ample anomaly over the Matusевич glacier, which was discovered during the aeromagnetic survey, was also examined gravimetrically in more detail. The main anomaly has its cause probably in larger depth of some kilometres.

In the **Bouguer anomaly** map (gravity survey at the Oates Coast during GANOVEX VIII; data source: BGR) and 3-D representation (Oates Coast during GANOVEX VIII; data source: BGR), one can observe the usual increase in gravity towards the coast due to the thinning of the Earth's crust. In addition, gravity increases clearly along the coast towards the west, obviously, like around 158/159°E, and then once more around 155°. It can be seen that a changing crustal thickness is related to the variability in gravity anomalies that can be checked by isostatic models. The large glaciers Rennick and Lillie are associated with strong negative anomalies in Bouguer estimates because of the mass deficits due to the lighter ice masses. It obviously possesses a reduced density but increased magnetic susceptibility. A strong positive anomaly, however, raises the block of the Bowers Mountains between the two glaciers mentioned. The figure (gravity survey at the Oates Coast during GANOVEX VIII; data source: BGR) represents the gravity and magnetic modelling obtained by joint inversion along a profile across the Matusевич glacier for better understanding of the interpreted results of the Oates Coast of Antarctica.

15.8.7 MARINE GRAVITY SURVEY OF MARINE FIELD: CASE STUDY FOR SILURIAN REEF EXPLORATION

A gravity survey conducted over and around Marine Field in south-western Illinois has been used as an example to show how measurement of the local gravity field can aid in the search for Silurian reefs in the Illinois basin (Heigold and Stephen, 1989). Acquisition parameters for gravity surveys over Silurian reefs should be calculated beforehand from simple models of the reef based on estimates of density contrasts, depths, and size. Gravity mapping may show an anomaly that coincides with a reef, providing the reef formed over a substantial enough basement high or has a strong density contrast with neighbouring inter-reef sediments. An example of a simple gravity anomaly over the reef at Marine Field (T 4N-R6W, Madison County) is shown in the figure in Heigold and Stephen (1989). The Bouguer gravity anomaly map shows anomaly at Marine Reef (T4N-R6W). Bouguer anomaly maps and derivative maps of gravity data can be effective methods for reef exploration in certain cases. Such an anomaly was partly responsible for the discovery of the reef at New Memphis Field (Bristol, 1974). Gravity data may also prove to be useful in locating eroded remnants of reefs. Not all reefs display gravity anomalies, however; nor does every gravity anomaly indicate the presence of a reef. Heigold and Stephen (1989) indicated that residual and derivative mapping techniques are power approaches for the enhancement of gravity anomalies for better matching with geological structures and relief on buried reefs. Interestingly, in the gravity survey, the role of the second vertical derivative map of the residual Bouguer gravity anomaly has special significance that allows a distinct comparison of marine-based gravity field data with the structure of the reef as mapped from subsurface data.

This study indicates that similar mapping techniques could be effective on other buried pinnacle reefs elsewhere in the world. Several sets of successful applications of the gravity surveys showed that it can be used either for reconnaissance survey as an initial method of exploration because of inherent density contrasts present within the different strata beneath the surface or can be used confidently in the exploration of prospective geological formation, suggesting further deployment of more extensive exploration techniques for smaller prospective areas as, for example, reef-prone zones.

15.8.8 MARINE MAGNETIC SURVEY OF A SUBMERGED ROMAN HARBOUR, CAESAREA MARITIMA, ISRAEL

Bowens (2009) and Boyce et al. (2002) showed that applications of marine magnetic surveys can range from archaeological mapping of shipwrecks or flooded archaeological sites to mapping of geological structures along with the identification of contaminant. The archaeological investigations of a historic harbour off the coast of Israel, completed by Boyce et al. (2002, 2004), is presented with survey coverage and resulting data. Resulting data from marine magnetic surveys, like surface surveys, are generally presented in 2-D contoured plan-view images ('Environmental geophysics—marine magnetic methods', Boyce et al., 2002 or Boyce and Reinhardt, 2004). The bathymetry data are corroborated with the total magnetic field data since there is enough evidence of correlating. The magnetic responses correlate to historic structures and geologic formations, shedding light on the pattern of magnetic anomalies along the specific geological structures and compositions of geologic materials.

15.8.9 UNEXPLODED ORDNANCE LOCATED BY MARINE MAGNETIC SURVEY, TOBRUK, LIBYA

Since 1960, integrated geophysical surveys by deploying both marine magnetometer and marine mounted seismic sensors have been deployed for high-resolution marine geophysical surveys. The detection of ferromagnetic objects on or near the seafloor along a survey vessel's track has become possible through the measurement of anomalies to the Earth's total magnetic field caused by the object. Typical features detected include sunken ships, pipelines, communication cables, anchors and chains, unexploded ordnance (Figure 15.5), metallic debris, and many other objects that could hinder the installation of structures on the seafloor. Magnetic anomalies are also associated with geological features, such as faults, volcanic intrusions, and other structures that produce lateral variations of the magnetic susceptibility contrasts of underlying sediments and rocks. This technique has been used extensively in the exploration for minerals and hydrocarbons.

15.8.10 BURIED VOLCANIC STRUCTURES IN THE GULF OF NAPLES (SOUTHERN TYRRHENIAN SEA, ITALY) RESULTING FROM HIGH-RESOLUTION MAGNETIC SURVEY AND SEISMIC PROFILING

A classical case study of marine-based high-resolution magnetic survey has been cited for the buried volcanic structures in the Gulf of Naples (Southern Tyrrhenian Sea, Italy), where high-resolution seismic profiling has also been carried out for the support of magnetic observations. Aiello and Marsella (2013) prepared a total magnetic field map for the Soma–Vesuvius volcanic complex in the Gulf of Naples associated with the most prominent lineaments and furnished the structural interpretation modified after Cassano and La Torre (1987), where they showed the trend of volcanites in the volcanic complex's peripheral areas. The new interpretation by Aiello and Marsella (2013) highlighted the fact that a subcircular anomaly is centred on the volcano, with two positive appendages diverging towards SE and SW. They might correspond to a great thickness of lava products, possibly in pre-existing depressions of the sedimentary basement of the graben of the Campania Plain, and suggests the occurrence of an elongated magnetized body that tends to move towards the Naples Bay from the Vesuvius volcano towards Torre del Greco. Moreover, Aiello and Marsella (2013) provided an alternative explanation for the presence of a strip of eruptive vents settled on a system of NE–SW normal faults. The beauty of a detailed magnetic map prepared by Mirabile et al. (2000) where data using sea beam



FIGURE 15.5 Unexploded ordnance located by marine magnetometer survey, Tobruk, Libya.

Source: Courtesy of Jack Hudson, Cultural Resource Services Inc., Seabrook, Tex.

bathymetry and magnetic surveys in a sector of the Naples Bay has been acquired, and they suggested there exist unknown underwater volcanoes (Mirabile et al., 2000). Moreover, these authors have evidenced the structure of the carbonate basement characterized by structural highs and tilted blocks at increasing depths. The strong down-throwing of the Phlegrean Fields area is confirmed by the seismic markers referred to the top of the Miocene unconformity and to the top of carbonate platforms.

The figure in Aiello et al. (2005) along the profile in a wide volcanic structure located offshore of the Torre del Greco town is related to the Somma–Vesuvius volcanic complex, showing a good correlation between volcanic structures and magnetic anomalies in the Gulf of Naples (Southern Tyrrhenian Sea) based on high-resolution magnetic profiling carried out during oceanographic cruise GMS00-05 (Aiello et al., 2005). The active volcanic area of the Late Quaternary in the Gulf of Naples has been mapped by Aiello et al. (2005) during the oceanographic cruise by deploying a densely spaced grid of magnetic profiles coupled with multichannel seismic method during the survey. They inferred that the correlations between seismic and magnetic profiles allow us to better assess the geological structure of the Gulf of Naples. Since Aiello et al. (2005) recognized shallow volcanic structures in the subsurface of the gulf by conducting seismo-stratigraphic analysis of high-resolution profiles, they arrived at the conclusion that the volcanic nature of some of these structures can be detected by identifying the magnetic anomalies on a high-resolution magnetic anomaly map of the gulf, which demonstrates the efficacy of the marine magnetic surveys in detection of volcanogenic anomalies.

REFERENCES

- Aiello, G., Angelino, A., D’Argenio, B., Marsella, E., Pelosi, N., Ruggieri, S. and Siniscalchi, A., 2005. Buried volcanic structures in the Gulf of Naples (Southern Tyrrhenian Sea, Italy) resulting from high resolution magnetic survey and seismic profiling. *Annals of Geophysics*, vol. 48, no. 6, pp. 883–897.
- Aiello, G. and Marsella, E., 2013. Integrated marine magnetics of the Naples Bay—from old to new data: Examples from Naples and Pozzuoli Gulfs (Southern Italy). GNGTS 2013, Sessione 3.2, pp. 86–91.
- Aiello, G., Marsella, E. and Ruggieri, S., 2010. Three-dimensional magneto-seismic reconstruction of the “Torre del Greco” submerged volcanic structure (Naples Bay, Southern Tyrrhenian sea, Italy): Implications for Vesuvius’ marine geophysics and volcanology. *Near Surface Geophysics*, vol. 8, no. 1, pp. 17–31.
- Atmanand, M.A., Purnima, J., Ramanamurthy, M.V., Ramadass, G.A., Ramesh, S., Gopakumar, K.A., Vedachalam, N. and Dharani G., 2019. Blue economy—opportunities for India. *IEEE India Info*, vol. 14, no. 3, pp. 106–115.
- Benetti, S., Dunlop, P. and Cofaigh, C.O., 2010. Glacial and glacially-related features on the continental margin of northwest Ireland mapped from marine geophysical data. *Journal of Maps*, pp. 14–29.
- Bowens, A., 2009. Geophysical and remote-sensing surveys. *The International Journal of Nautical Archaeology*, vol. 2, no. 13, pp. 103–113.

- Boyce, J.I., Pozza, M.R., Morris, W.A., Eyles, N. and Doughty, M., 2002. High-resolution magnetic and seismic imaging of basement faults in Western Lake Ontario. Conference Proceedings, University of Toronto and McMaster University.
- Boyce, J.I. and Reinhardt, E.G., 2004. Marine magnetic survey of a submerged Roman Harbour, Caesarea Maritima, Israel. *The International Journal of Nautical Archaeology*, vol. 33, no. 1, pp. 122–136, DOI: 10.1111/j.1095-9270.2004.010.x.
- Boyce, J.I., Reinhardt, E.G., Raban, A. and Pozza, M.R., 2004. Marine magnetic survey of a submerged Roman Harbour, Caesarea Maritima, Israel. *The International Journal of Nautical Archaeology*, vol. 33, no. 1, pp. 122–136.
- Bristol, H.M., 1974. Silurian pinnacle reefs and related oil production in southern Illinois: Illinois state geological survey. *Illinois petroleum*, vol. 102, p. 97.
- Camidge, K., Holt, P., Johns, C., Randall, L. and Schmidt, A., 2010. Developing magnetometer techniques to identify submerged archaeological sites—theoretical study report. Cornwall Council, Historic Environment Projects Rep 2010R012. Kennall Kennall Building, Old County Hall, Station Road, Truro, Cornwall, UK, TR1 3AY. <https://www.3hconsulting.com/Downloads/Developing%20Magnetometer%20Techniques%20Theoretical%20Study%20Final%20Report%20Rev%202.pdf>.
- Cassano, E. and La Torre, P., 1987. Geophysics. In: Santacroce, R. (ed), *Somma-Vesuvius*, Roma, Italy: CNR, Quaderni de la Ricerca Scientifica.
- Chandler, M.T. and Wessel, P., 2008. Improving the quality of marine geophysical track-line data: Along-track analysis. *J. Geophys. Res.*, vol. 113, B02102, DOI: 10.1029/2007JB005051. Wiley Online Library ADS Web of Science @ Google Scholar.
- Chandler, M.T. and Wessel, P., 2012. Errata-based correction of marine geophysical trackline data. *Published by AGU and Geochemical Society*, vol. 13, no. 10, pp. 1–17, DOI: 10.1029/2012GC004294.
- Cronin, B.T., Akhmetzhanov, A.M., Mazzini, A., Akhmanov, G., Ivanov, M. and Kenyon, N.H., 2005. Morphology, evolution and fill: Implications for sand and mud distribution in filling deep-water canyons and slope channel complexes. *Sedimentary Geology*, vol. 179, no. 1–2, pp. 71–97.
- Desa, M.A. and Ramana, M.V., 2021. Integrated analysis of magnetic, gravity, and multichannel seismic reflection data along a transect southeast of Sri Lanka, Bay of Bengal: New constraints. *Marine Geology*, Elsevier, vol. 438, p. 106543.
- Devas, V., Varghese, S., Kartikeyan, M., Singh, S.K., Mishra, R.K., Bara, A., Mohammad, M.P.A., Nandi, B.K. and Pathan, M.S., 2015. The morphological setup of Andaman-Nicobar trench and accretionary prism: A study using multibeam bathymetry. *Indian Journal of Geosciences*, vol. 69, no. 3–4, pp. 215–222.
- Drake, C.L., 1970. Marine geophysics. *Journal of American Geophysical Union (AGU)*, vol. 52, no. 6, pp. IUGG244–IUGG249.
- Fish, J.P. and Carr, H.A., 1990. *Sound Underwater Images: A Guide to the Generation and Interpretation of Side Scan Sonar Data*. Orleans, MA: Lower Cape Publishing.
- GSI, 2016. Institutional report. *Proc. Indian Natn. Sci. Acad.*, vol. 82, no. 3 July Spl Issue 2016, pp. 1061–1081. Printed in India, DOI: 10.16943/ptinsa/2016/48504.
- Heigold, P.C. and Stephen, T.W., 1989. A gravity survey of marine field case study for Silurian reef exploration. *Illinois Petroleum*, vol. 133.
- Hein, J.R., Mizell, K., Koschinsky, A. and Conrad, T.A., 2013. Deep-ocean mineral deposits as a source of critical metals for high- and green-technology applications: Comparison with land-based resources. *Ore Geology Reviews*, vol. 51, pp. 1–14.
- Kuhn, T., Węgorzewski, A., Rühlemann, C. and Vink, A., 2017. In *Deep-Sea Mining* (edited by R. Sharma). Cham, Switzerland: Springer International Publishing AG, pp. 23–63.
- Kumar, R.S. and Begum, S.K., 2022. Interpretation of gravity and magnetic data in the Central Indian Ocean. *Contributions to Geophysics and Geodesy*, vol. 52, no. 3, pp. 359–393.
- Mirabile, L., De Marinis, E. and Frattini, M., 2000. The phlegrean fields beneath the sea: The underwater volcanic district of Naples, Italy. *Bollettino di Geofisica Teorica Applicata*, vol. 41, no. 2, pp. 159–186.
- Mitchell, T. and Smith, K., 2014. Investigations of Marine Geohazards for Coastal Infrastructure Projects. *Sea Technology Magazine*, March.
- Moore, D.G. and Keller, G.H., 1984. Marine sediments, geotechnical properties. In: Finkl, C. (ed), *Applied Geology*. Encyclopedia of Earth Sciences Series, 3. Boston, MA: Springer, DOI: 10.1007/0-387-30842-3_41.
- Murthy, K.S.R., 2016. Baseline geophysical data for hazard management in coastal areas in relation to earthquakes and tsunamis. *The International Journal of Ocean and Climate Systems*, vol. 7, no. 1, pp. 27–31.
- Murthy, K.S.R., Subrahmanyam, A.S. and Subrahmanyam, V., 2012. *Tectonics of the Eastern Continental Margin of India*. New Delhi, India: The Energy and Resources Institute (TERI) Press, p. 180.

- Murthy, K.S.R., Subrahmanyam, V. and Murty, G.P.S., 2006. Impact of coastal morphology, structure and seismicity on the tsunami surge. In: Murty, T.S., Aswathanarayana, U., Nirupama, N. (eds), *The Indian Ocean Tsunami*. London: Taylor & Francis, pp. 19–31.
- Murthy, K.S.R., Subrahmanyam, A.S., Murty, G.P.S., et al. 2006a. Factors guiding tsunami surge at the NagapattinamCuddalore shelf, Tamil Nadu, east coast of India. *Current Science* vol. 90: 1535–1538.
- Murthy, K.S.R., Subrahmanyam, V., Murty, G.P.S., et al. 2006b. Impact of coastal morphology, structure and seismicity on the tsunami surge. In: Murty TS, Aswathanarayana U, Nirupama N (eds) *The Indian Ocean Tsunami*. London: Taylor & Francis, pp. 1931.
- Murthy, K.S.R., Subrahmanyam, V. and Subrahmanyam, A.S., 2010. Land-Ocean Tectonics (LOTs) and the associated seismic hazard over the Eastern Continental Margin of India (ECMI). *Natural Hazards*, vol. 55, pp. 167–175.
- Murthy, K.S.R., 2011. Some factors contributing to the changes in coastal morphology A case study from the Eastern Continental Margin of India. In: *Proceedings of international conference on 'impact of climate change on coastal ecosystem' (ICC ECO 2011)*, Sathyabama University, Chennai, India, 10–11 November.
- Murthy, K.S.R., Subrahmanyam, A.S., Subrahmanyam, V., 2012. *Tectonics of the Eastern Continental Margin of India*. New Delhi, India: The Energy and Resources Institute (TERI) Press, p. 180.
- Murthy, K.S.R., 2016. Baseline geophysical data for hazard management in coastal areas in relation to earthquakes and tsunamis, *Journal of Ocean and Climate: Science, Technology and Impacts*, <https://doi.org/10.1177/1759313115623161>.
- Murty, G.P.S., Subrahmanyam, A.S. and Murthy, K.S.R., 2002. Evidence of fault reactivation off Puducherry coast from marine geophysical data. *Current Science*, vol. 83, pp. 1446–1449.
- Nayak, S.R., 2023. Towards stewardship of oceans: Role of G20 countries. *Current Science*, vol. 124, no. 9, pp. 1019–1020.
- Nayak, S.R., 2024. From ocean science to sustainable blue economy. *Current Science*, vol. 126, no. 2, pp. 155–160, DOI: 10.18520/cs/v126/i2/155-160.
- O'Reilly, B.M., Shannon, P.M. and Readman, P.W., 2007. Shelf to slope sedimentation processes and the impact of Plio-Pleistocene glaciations in the northeast Atlantic, west of Ireland. *Marine Geology*, vol. 238, no. 1–4, pp. 21–44.
- Radhakrishna, M., Srinivasa Rao, G., Nayak, S., Bastia, R. and Twinkle, D., 2012. Early cretaceous fracture zones in the Bay of Bengal and their tectonic implications: Constraints from multi-channel seismic reflection and potential field data. *Tectonophysics*, vols. 522–523, pp. 187–197.
- Rajeevan, M., 2018. Why should India invest in deep ocean research? *Current Science*, vol. 115, no. 5, pp. 807–808.
- Rao, P.R., Meetei, L.I., Nandi, B.K., Pathan, M.S., Satyanarayana, G.V., Chakravarthi, D., Singh, R. and Nagabhusana Rao, J.A., 2016. Inferred seabed and sub-seabed palaeo-morphological features by shallow seismic investigations off north Tamil Nadu Coast, Bay of Bengal. 38th Annual Convention and Seminar on Exploration Geophysics, 20–22 October, Hyderabad.
- Ruth, P., Justin, D. and Richard, B., 2013. *Marine Geophysics Data Acquisition, Processing and Interpretation* (edited by D.M. Jones). Swindon: English Heritage Publishing, pp. 1–47.
- Satyanarayana, G.V., Nandi, B.K., Rao, P.R. and Pathan, M.S., 2015. Fluctuation of Curie isotherm depth around Barren Island, Andaman Sea. *Indian Journal of Geosciences*, vol. 69, no. 3, pp. 287–292.
- Sengupta, B.J., Ghatak, S.K., Nandi, B.K. and Pal, B.P., 2005. Analysis of temporal changes of magnetic (TF) anomaly around Barren and Narcondam Island and its significance in terms of the thermal state in the crust. *Journal of Geological Society of India*, vol. 65, pp. 97–100.
- Sharma, R. and Smith, S., 2019. In *Environmental Issues of Deep-Sea Mining* (edited by R. Sharma). Cham, Switzerland: Springer, pp. 3–22.
- Stoker, M.S., 1995. The influence of glacial sedimentation on slope apron development on the continental margin off Northwest Britain. *Geological Society, London, Special Publication*, vol. 90, pp. 159–177.
- Subrahmanyam, A.S., Murty, G.P.S. and Sarma, K.V.L.N.S., 2007. Qualitative assessment of tectonic lineaments over the coastal and inner shelf of Kakinada and Kalingapatnam, central east coast of India. *Journal of Geological Society of India*, vol. 69, pp. 1328–1334.
- Trabant, P.K., 1984. *Applied High Resolution Geophysical Methods*. Dordrecht: Springer Science Business Media. DOI: https://doi.org/10.1007/978-94-009-6493-8_1
- Wadhawan, S., 2016. Marine geology and coastal surveys for seabed mapping and mineral resource evaluation in India—an overview. *Proceedings of the Indian National Science Academy*, vol. 82, no. 3, pp. 675–683.
- Wessel, P. and Chandler, M.T., 2011. The spatial and temporal distribution of marine geophysical surveys. *Journal of Geophysical Research*, United States, vol. 59, no. 1, pp. 55–71.

16 A Comparative Overview of Geophysical Methods with Equation Dictionaries, Field Layouts, and Instrumental Setups

The advent of geophysical sciences steered the geoscientific community to work for prosperity with the sustainability of natural subsurface resources, and such endeavour of the last two decades directed us to estimate the shallow as well as deeper subsurface structures of the Earth, which is important to understand for many economic and safety reasons. The problem is usually difficult due to the complexity of the subsurface processes, especially near the surface. Based on physical property contrast, a number of geophysical methods are used to characterize the Earth's materials. The geophysical method allows us to understand the problem and propose a plausible but reliable solution in accordance with the constraint of the dataset used for different methods. In order to synthesize information from different geophysical methods, it is important to understand their similarities and dissimilarities. The chapter aims to correlate the basic principles of geophysical methods side by side, starting from the fundamental equations. The geophysical study reveals that there exist many analogies among different methods with reference to both in their mathematical formulation and, sometimes, in the ways they are used in geophysical applications.

16.1 OVERVIEW OF GEOPHYSICAL METHODS

Geophysics is a branch of the Earth sciences which aims to understand the Earth using the methods of physics. As mentioned previously, the geophysical application areas are generally related to our understanding of the subsurface structure of the Earth for both scientific and economic reasons. A general classification of the methods currently used in the geophysical arena and some of their basic characteristics are given in Table 16.1.

16.2 FUNDAMENTAL PROPERTIES OF SOURCES

Based on the fundamental and conceptual physics having important role of the Newtonian physics, all the forces that can act on a body with mass m , electric charge q , and velocity v can be stated by the Lorentz equation with the addition of the gravitational term as

$$F = m \cdot g + q \cdot E + q \cdot v \times B$$

where $g = (g_x, g_y, g_z)$, $E = (E_x, E_y, E_z)$, and $B = (B_x, B_y, B_z)$.

In this equation, the vectors g , E , B are called the force fields. It is evident that E and B are fundamentally related to each other, as they both act on the same charge q . In fact, although theoretically natural magnetic monopoles can exist, they have not been observed so far; so all the magnetic fields observed in nature are assumed to be created by electric sources. As opposed to gravity, which

TABLE 16.1
Summary of Geophysical Methods and Their Characteristics Applicable to the Exploration and Geoenvironmental Studies

Method	Physical Parameter Measured	Typical Units	Relevant Physical Property (Investigated)	Typical Source of Anomaly	Depth of Investigation
Gravity: A, B, G	Total attraction of Earth's gravity field (the vertical attraction of anomalous masses)	Milligals or gravity unit (0.1 mGal) Eötvös unit (10 gal/cm)	Density	Rock density contrasts	All
Magnetic: A, B, G	Gradient of Earth's gravity field				
	Vector component, or total attraction of Earth's magnetic field	Nanotesla or gammas Nanotesla/m	Magnetic susceptibility and remanent magnetization	Magnetic susceptibility and/or remanent magnetization contrasts	Surface to Curie isotherm
	Gradient of Earth's magnetic field				
Gamma ray scintillometry: A, B, G	Rate of gamma ray photons received	Counts/sec Counts/sec in spectral regions; if calibrated, %K and ppm equiv. U and Th	Quantity of K, U, Th and daughters	K, U, and Th contrasts in Earth's upper 50 cm	Upper 50 cm
Gamma ray spectrometry: A, B, G	Rate of gamma ray photons received and their energy		Quantity of K, U, Th and daughters	K, U, and Th contrasts in Earth's upper 50 cm	
Seismic refraction: B, G	Seismic energy travel time	Milliseconds	Velocity of P or S waves	Structures or velocity layer contrasts	All
Seismic reflection: B, G					
Thermal borehole or shallow hole: B	Thermal gradient or temperature	°C/m, °C	Thermal conductivity	Thermal flux or conductivity variations	Hole depth
Thermal remote sensing: A, G	Surface temperature, day and night	°C	Thermal inertia	Thermal inertia contrasts	About 5 cm
Electrical direct current resistivity: B, G	Electrode position (m), applied current (A), and electric field (mV)	Metre, amps, millivolts; typically converted to units of resistivity (Ω -m)	Resistivity	Lateral or vertical changes in resistivity	About 2 km
Several variations in electrode geometry					
Electromagnetic methods: A, B, G	Dependent on method; ratio of received to applied electric and magnetic fields	Impedance (ohms) or dimensionless ratio; units of conductivity (siemens/m) or resistivity (Ω -m)	Conductivity (inverse of resistivity)	Lateral or vertical changes in Earth conductivity	Shallow (10 m, VLF; 100 m, controlled source), intermediate (1 km, AMT), deep (10 km, MT)
Many variations available					
Mise-a-la-masse: B, G	Applied DC or low-frequency AC field	Millivolts	Resistivity	Conductive body	A few hundred metres

(Continued)

TABLE 16.1 (Continued)
Summary of Geophysical Methods and Their Characteristics Applicable to the Exploration and Geoenvironmental Studies

Method	Physical Parameter Measured	Typical Units	Relevant Physical Property (Investigated)	Typical Source of Anomaly	Depth of Investigation
Induced polarization: B, G	Resistivity change/frequency (PFE)	Percent change	Interface ionic polarization	Metallic lustre minerals and pore water Clay and zeolite minerals	About 2 km
	Phase angle between transmitted and received signal (ϕ)	Milliradians			
	Normalized area of part of received voltage decay curve	Milliseconds			
Self-potential: B, G	Natural near-static (direct current) electric field	Millivolts	Eh/pH electronic conductor; streaming potential and thermal coupling coefficients	Vertical change in Eh/pH caused by electronic conductor; groundwater flow; thermal flux	A few hundred metres
Remote sensing: A	Reflected radiation intensity (UV, VIS, IR)	Recorded as optical or digital intensity image	Spectral reflectance, albedo	Changes in spectral reflectance and albedo	Surface only

Note: A: airborne surveys; B: borehole surveys; and G: ground surveys.

TABLE 16.2
Basic Properties of Sources of Force Fields

Field	Type	Polarization	Conduction
Gravity	Scalar potential field	No	No
Magnetic	Vector	Yes	No
Electric	Scalar/vector	Yes	Yes

is represented by monopoles, electric and magnetic sources can have both positive and negative charges (Table 16.2). The interaction of electric and magnetic fields inside a medium with sources becomes complicated as they polarize the medium passing through.

16.3 SOME MATHEMATICAL IDENTITIES

1. $\nabla = i \frac{\partial}{\partial x} + j \frac{\partial}{\partial y} + k \frac{\partial}{\partial z}$, where i, j, k are three unit vectors.
2. $\nabla \times (\nabla U) = 0$, that is, the curl of the gradient is zero, and U is a scalar function.
3. $\nabla \cdot (\nabla \times \mathbf{F}) = 0$, that is, the divergence of the curl is zero.
4. $\nabla^2 \mathbf{F} = \nabla (\nabla \cdot \mathbf{F}) - \nabla \times (\nabla \times \mathbf{F})$ (Laplacian of a vector field).
5. $\nabla^2 U = \nabla \cdot (\nabla U)$ (Laplacian of a scalar field).
6. $\nabla \times \mathbf{E} = \begin{vmatrix} i & j & k \\ \frac{\partial}{\partial x} & \frac{\partial}{\partial y} & \frac{\partial}{\partial z} \\ E_x & E_y & E_z \end{vmatrix}$ and $\nabla \times \mathbf{H} = \begin{vmatrix} i & j & k \\ \frac{\partial}{\partial x} & \frac{\partial}{\partial y} & \frac{\partial}{\partial z} \\ H_x & H_y & H_z \end{vmatrix}$.

16.4 EQUATION DICTIONARIES IN INTRODUCTORY GEOPHYSICS

The creation of the equation dictionary of introductory geophysics is best written in table form, as follows.

16.4.1 GRAVITY

BASIC PRINCIPLES

Newton's law of gravitation:

$$F = -G \frac{m_1 m_2}{r^2} \mathbf{r}_1$$

where F = force in dynes, m_1 and m_2 masses in gms; r = distance in cm; G = gravitational constant = 6.67×10^{-8} dyne-cm²/gm²; $\mathbf{r}_1$ = unit vector.

Acceleration of gravity (g) due to a point mass (m) a distance (r):

$$g = \frac{Gm}{r^2}$$

In SI system, $g = 9.8 \text{ m/sec}^2$.

Mass of the Earth

$$M = \frac{gR^2}{G}$$

where R = radius of the Earth = 6,400 km; $G = 6.67 \times 10^{-8} \text{ c.g.s. unit}$; $g = 981 \text{ cm/sec}^2$; we get $M = 6.1 \times 10^{27} \text{ gm}$.

Gravitational Potential

$$U = -\frac{Gm}{r}$$

The first derivative of U in any direction gives the value of gravity in that direction:

$$g_z = \frac{\partial U}{\partial z}$$

Potential Field Equations

The Gauss's divergence theorem:

$$\int \nabla \cdot g \, dV = \int g_n \cdot dS$$

Laplace equation for gravitational field in free space:

$$\nabla^2 U = 0$$

Poisson's equation for the gravitational field in an attracting medium:

$$\nabla^2 U = 4\pi G\sigma$$

where σ = density.

Gradient of Gravity

$$\frac{\partial g_z}{\partial z} = \frac{\partial^2 U}{\partial z^2}$$

Second Derivative

$$\frac{\partial^2 g_z}{\partial z^2} = \frac{\partial^3 U}{\partial z^3}$$

Shape of the Earth

Clairaut's theorem, which states:

$$\frac{R_e - R_p}{R_e} + \frac{g_p - g_e}{g_e} = \frac{5}{2} \frac{R_e \omega^2}{g_e}$$

ω = angular frequency; R_e = equatorial radius = 6,378 km; R_p = 6,357 km; $R_e\omega^2$ = the centrifugal acceleration; f = geometrical flattening = $\frac{R_e - R_p}{R_e} = \frac{1}{298.252}$; *gravitational flattening* = $\frac{g_p - g_e}{g_e}$.

The radius of the sphere:

$$R = \left(R_e^2 R_p \right)^{\frac{1}{3}}$$

Spheroid and Geoid

Clairaut's gravity formula relates gravity to latitude on the reference spheroid according to an equation of the form

$$g_\varphi = g_0 (1 + k_1 \sin^2 \varphi + k_2 \sin^2 2\varphi)$$

According to the 1930 formula:

$$g_\varphi = 978.049 (1 + 0.0052884 \sin^2 \varphi - 0.0000059 \sin^2 2\varphi) \text{ Gal}$$

According to the 1967 formula:

$$g_\varphi = 978.0318 (1 + 0.0053024 \sin^2 \varphi - 0.0000059 \sin^2 2\varphi) \text{ Gal}$$

$$g_{1967} - g_{1930} = (-17.2 + 13.6 \sin^2 \varphi) \text{ mGal}$$

$$g_{1980} - g_{1930} = (-16.3 + 13.7 \sin^2 \varphi) \text{ mGal}$$

where g_φ = predicted value of gravity at a latitude (φ); g_0 = value of gravity at the equator; and k_1 and k_2 are constants, dependent on the shape and speed of rotation of the Earth. However, the gravity relation given earlier is only approximate because of warping in the equipotential surface of gravity. The actual equipotential surface of the Earth that corresponds to sea level is called the geoid.

Variation of g from Place to Place

Variation in acceleration due to gravity due to altitude:

$$g = \frac{GM}{(R + h)^2}$$

where h = height from the surface of the Earth; R = radius of the Earth; M = mass of the Earth.

Variation in acceleration due to gravity due to depth:

$$g = \frac{4}{3} G\pi (R - d) \rho$$

where d = depth from the surface of the Earth, ρ = density of the Earth.

MEASUREMENT OF GRAVITY ON LAND

Measurement of Absolute Gravity

The period of a physical pendulum may be written as

$$T = 2\pi \sqrt{\frac{I}{mgh}}$$

where I is the moment of inertia about the knife edge, m is the total mass of the pendulum, h is the distance from knife-edge to the centre of gravity, and g is the acceleration due to gravity.

In the Kater's reversible pendulum, the factor $\frac{I}{mh}$ is replaced by L , the length of equivalent simple pendulum. Therefore:

$$T = 2\pi \sqrt{\frac{L}{g}}$$

The sensitivity of the pendulum:

$$\frac{dg}{g} = -2 \frac{dT}{T}$$

Measurement of Relative Gravity

According to Hooke's law:

$$F = mg = -k(s - s_0)$$

where m = mass, s_0 = spring length, s = new length, k = spring constant, F = force, g = acceleration due to gravity.

The sensitivity of unstable or astatized types of gravimeter:

$$\frac{\delta l}{\delta g} = \frac{m}{k}$$

where a spring of length (l) has been stretched to $(l + \delta l)$ as a result of increase of gravity (dg), increasing the weight of the suspended mass (m). If k is reduced to $(k - k')$ by the application of some force (astatizing force or labializing force), we get an astatic gravimeter. Then:

$$T = \frac{2\pi m}{k - k'}$$

where k' is the constant for labializing force. It is clear that the closer the value of k' approaches k , the greater the sensitivity.

Sensitivity, Accuracy, and Resolution of Gravimeter

The sensitivity of modern gravimeter Scintrex CG-5 Autograv on land is 0.01 mGal, accuracy 0.1 mGal, resolution 0.001 mGal (1 μ Gal). Sensor-type fused quartz using electrostatic nulling.

Measurement of Component of Gravity

$$dg = dg_z$$

GRAVITY SURVEYING

Reduction of Gravity Data

Generalized summary of the types of corrections that we have applied to the gravity data, drift correction, tidal correction, latitude correction, elevation correction, Eötvös correction, and isostatic correction.

Latitude Correction

$$g_{\varphi} = 978.0327(1 + 0.0052790414 \sin^2\varphi - 0.0000232718 \sin^4\varphi + 0.0000001262 \sin^6\varphi) \text{ Gal.}$$

Elevation correction (free air correction [FAC], Bouguer correction [BC], and terrain correction [TC]):

Free air correction (FAC) = $0.3086 h$ mGal/m, where h is height in m from sea level.

Bouguer correction (BC) = $2\pi G \rho h = 0.04191 \rho h$ mGal/m, where ρ is density in gm/cc.

Combined elevation correction factor = $(0.3086 - 0.04191 \times 2.67) h$ mGal/m = $0.1967 h$ mGal/m.

Terrain Correction

$$\Delta g_T = G \rho \varphi_0 \left[\left(\sqrt{r^2 + h^2} - r_1 \right) - \left(\sqrt{r^2 + h^2} - r_2 \right) \right]$$

Eötvös Correction

$$E_C = 7.503 V \cos\varphi \sin\alpha + 0.004154 V^2$$

where V is in knots, φ is the latitude, α is the direction of movement w.r.t. the north.

GRAVITY ANOMALIES

Free Air Anomaly (FAA)

$$\text{FAA} = g_{\text{obs}} + \text{FAC} - g_{\varphi}$$

Bouguer Anomaly (BA)

$$\text{BA} = g_{\text{obs}} + \text{FAC} (-, +) \text{BC} + \text{TC} - g_{\varphi}$$

where observed gravity (g_{obs}) is the gravity readings observed at each gravity station after corrections have been applied for instrument drift and Earth tides.

Isostatic Anomaly (IA)

$$\text{IA} = g_{\text{obs}} + \text{FAC} - \text{BC} + \text{TC} + \text{isostatic correction} - g_{\varphi}$$

ISOSTASY

Airy hypothesis:

$$r = \frac{h \rho_c}{\rho_m - \rho_c}$$

where ρ_c = crustal density, ρ_m = mantle density, r = root depth, h = mountain height.

ROCK DENSITY

$$\rho_r = \frac{W_a}{W_a - W_w}$$

ρ_r = density of rock sample relative to water, w_a = weight in air, w_w = weight in water.

INTERPRETATION OF GRAVITY ANOMALIES

The gravity anomaly of a sphere:

$$\Delta g_z = \frac{4}{3} \pi G \left(\frac{\Delta \rho R^3}{z^2} \right) \frac{1}{\left[1 + \left(\frac{x}{z} \right)^2 \right]^{\frac{3}{2}}}$$

where R = radius of the sphere, $\Delta \rho$ = density contrast, x = horizontal distance from sphere.

The depth to the centre of the sphere:

$$z = 0.652 w$$

The width (w) of the anomaly where the amplitude has one-half its maximum value is called the 'half-height width'.

The gravity anomaly of a horizontal line element:

$$\Delta g_z = \frac{2Gmz}{z^2 + x^2}$$

where infinitely long linear mass distribution with mass m per unit length extends horizontally along the y axis at depth z .

The gravity anomaly of a horizontal cylinder:

$$\Delta g_z = 2\pi G \left(\frac{\Delta \rho R^2}{z} \right) \left[\frac{1}{1 + \left(\frac{x}{z} \right)^2} \right]$$

where R = radius of the cylinder, $\Delta \rho$ = density contrast, x = horizontal distance from cylinder.

The depth to the axis of the cylinder:

$$z = 0.5 w.$$

The gravity anomaly of horizontal thin sheet:

$$\Delta g_z = 2G \Delta \rho t \left[\frac{\pi}{2} + \tan^{-1} \left(\frac{x}{z} \right) \right]$$

where t = thickness of the sheet.

The gravity anomaly of an infinite horizontal sheet:

$$\Delta g_z = 2\pi G \Delta \rho t$$

It extends to infinity in the positive and negative x and y directions and same as the expression for the Bouguer plate correction.

SOLAR SYSTEM DATA

Polar radius of the Earth is 6,357 km.

Equatorial radius of the Earth is 6,378 km.

Increase of gravity from equator to pole is roughly 5,186 mGal.

Mass of the Earth: $M_E = 5.97 \times 10^{24}$ kg.

Mass of the sun: $M_S = 1.99 \times 10^{30}$ kg.

Mass of the moon: $M_M = 7.36 \times 10^{22}$ kg.

Earth–moon distance: 3.84×10^5 km.

Earth–sun distance: 1.50×10^8 km.

16.4.2 MAGNETIC

MAGNETIC MONOPOLES (COULOMB'S LAW)

The mathematical expression for the magnetic force experienced between two magnetic monopoles is given by

$$F_m = \frac{1}{\mu} \frac{p_1 p_2}{r^2}$$

where μ = magnetic permeability = $4\pi \times 10^{-7}$ H/m; p_1 and p_2 are the strengths of the two magnetic monopoles; and r is the distance between the two poles.

Units Associated with Magnetic Poles

Magnetic field strength:

$$H = \frac{F_m}{p_2} = \frac{p_1}{\mu r^2} \text{ in nT or gamma}$$

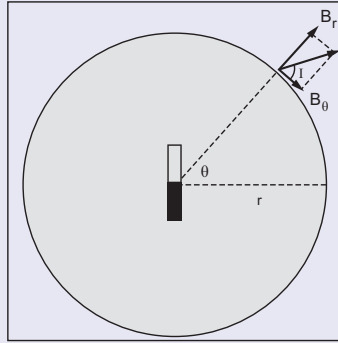
The average strength of the Earth's magnetic field is about 50,000 nT.

Magnetic (dipole) moment:

$$M = IA \text{ in A m}^2$$

where I is the intensity of magnetization and A is the area. The direction of M is perpendicular to the area of the current loop.

DIPOLAR MAGNETIC FIELD



$$B_r = \frac{2M\mu_0\sin\theta}{4\pi r^3} \text{ and } B_\theta = \frac{M\mu_0\cos\theta}{4\pi r^3}$$

where M = magnetic moment, μ_0 = magnetic permeability = $4\pi \times 10^{-7}$ H/m.

$$\tan I = 2\tan\theta$$

$$F(r, \theta) = \sqrt{B_r^2 + B_\theta^2} = \frac{M\mu_0\sqrt{3\sin^2\theta + 1}}{4\pi r^3}$$

MAGNETIZATION OF MATERIALS

Magnetic Induction or Flux

$$B = \mu H$$

Magnetic Susceptibility

$$k = \frac{I}{H}$$

Susceptibility k , 1 cgs unit = $\frac{1}{4\pi}$ SI unit; I = intensity of magnetization, 1 cgs unit (emu) = 10^3 A / m in SI unit.

Maximum Induced Magnetic Field Strength (i.e. the Maximum Anomaly)

For a 3-D buried object:

$$T = \frac{M}{r^3}$$

For latitudes greater than 60° :

$$T = \frac{2M}{r^3}$$

For a 2-D buried object:

$$T = \frac{M}{r^2}$$

where r = distance to the target; M = magnetic moment = kHV ; V = volume. The magnetic properties of most rocks depend on the amount of the magnetite (Fe_3O_4) within the rocks.

Average susceptibilities for specific magnetic minerals (in SI unit):

Magnetite (Fe_3O_4) = $6,000 \times 10^{-3}$; ilmenite (FeTiO_3) = $1,800 \times 10^{-3}$; hematite (Fe_2O_3) = 6.5×10^{-3} ; pyrite (FeS_2) = 1.5×10^{-3} ; pyrrhotite (Fe_7S_8) = $1,500 \times 10^{-3}$.

MAXWELL EQUATION RELATED TO MAGNETISM

Gauss's law for magnetism:

$$\nabla \cdot \mathbf{B} = 0$$

By the divergence, this gives

$$\oiint \mathbf{B} \cdot d\mathbf{A} = 0$$

That is, no net sources or sinks of magnetic flux, or no magnetic point charges or monopoles. Unlike Gauss's law for gravity:

$$\oiint \mathbf{g} \cdot d\mathbf{A} = 4\pi G M$$

Ampere's law for magnetism:

$$\nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t}$$

where $\mathbf{J}$ = current density in vector, $\mathbf{D}$ = electric displacement in vector. For static magnetic field in the air, $\mathbf{J} = 0$ and $\frac{\partial \mathbf{D}}{\partial t} = 0$, that is, no current, no changing electric fields ($\nabla \times \mathbf{H} = 0$).

By Stokes theorem:

$$\oint \mathbf{H} \cdot d\mathbf{l} = \oiint \nabla \times \mathbf{H} \cdot d\mathbf{A} = 0$$

Because the line integral around a closed loop = 0, the magnetic field intensity is a conservative field.

MAGNETIC POTENTIAL

Magnetic potential (U):

$$\mathbf{H} = -\nabla U$$

Since magnetic field intensity $\mathbf{H}$ is a conservative field. Note that this is an identical formulation to gravity:

$$\mathbf{g} = \nabla U$$

where U is gravitational potential.

MAGNETIC FIELD NOMENCLATURE

Declination (D) is the angle between the magnetic north and the horizontal projection of the magnetic field ($\mathbf{F}$):

$$\tan D = \frac{Y}{X}$$

X and Y are components of H (horizontal component of F) to the north and east, respectively.

Inclination (I) is the angle between the surface of the Earth and F :

$$\sin I = \frac{Z}{F}$$

where Z = vertical component of F positive downward.

TOTAL MAGNETIC FIELD MEASUREMENTS

Total magnetic field:

$$F_t = F_e + F_a$$

where F_e = Earth's main magnetic field, F_a = anomalous magnetic field (induced and remanent magnetization).

CORRECTIONS FOR TEMPORAL VARIATIONS OF EARTH'S MAGNETIC FIELD

The magnetic field varies with time like the gravitational field. Corrections are adopted for three temporal variations:

1. *Secular variations*. These are long-term (changes in the field that occur over years) variations of the Earth's main magnetic field.
2. *Diurnal variations*. These are variations in the magnetic field that occur over the course of a day (20–30 nT per day).
3. *Magnetic storms*. Occasionally, magnetic activity in the ionosphere will abruptly increase (amplitude changes as large as 1,000 nT).

INSTRUMENTS

There are two main types of ground magnetometer in common survey use for mineral exploration, namely, the fluxgate and proton-free precession magnetometer (Breiner, 1973).

Fluxgate magnetometer MFD-4: range $\pm 100,000$ nT, sensitivity 10 or 20 nT. However, Scintrex manufactures a 1 nT digital fluxgate magnetometer, the MFD-4, in which the vertical component is read from a five-digit light-emitting diode display. Fluxgate gradiometer has a differential resolution of 0.5 nT.

GEM GSM-19T proton precession magnetometer: sensitivity 0.15 nT, resolution 0.01 nT, absolute accuracy ± 0.2 nT.

16.4.3 SEISMOLOGY

SEISMIC WAVES

Stress and Strain

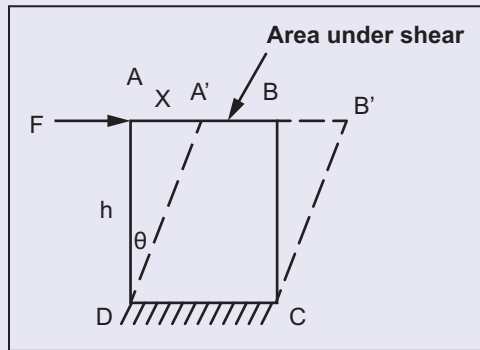
$$\text{Stress} = \frac{F}{A}$$

where F is the force and A is cross-sectional area. SI unit of stress is pascal $\left(1\text{Pa} = \frac{\text{N}}{\text{m}^2}\right)$.

$$\text{Strain} = \frac{\Delta h}{h}$$

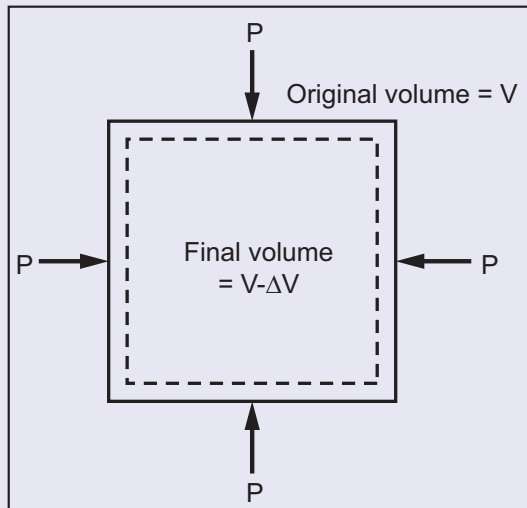
Fractional change in dimension, which is a dimensionless quantity; h = height of a right cylindrical block. For elastic behaviour, the strain in a body is proportional to the stress applied to it. This linear relationship $\left(\frac{F}{A} \propto \frac{\Delta h}{h}\right)$ is called Hooke's law. It forms the basis of the elasticity theory.

$$\text{Shear stress, } P = \frac{F}{A}; \text{ shear strain, } \tan\theta = \theta = \frac{x}{h}$$



$$\text{Shear modulus} = \text{modulus of rigidity } (\mu) = \frac{\text{shear stress}}{\text{shear strain}} = \frac{Fh}{Ax}$$

$$\text{Volume stress} = P, \text{ volume strain} = \frac{\Delta V}{V}$$



$$\text{Bulk modulus } (K) = \frac{\text{volume stress}}{\text{volume strain}} = \frac{P}{\frac{\Delta V}{V}}$$

Types of Seismic Waves

1. Body waves (*P* wave, *S* wave)
2. Surface waves (Rayleigh wave, Love wave)

SEISMIC WAVE VELOCITIES

$$V_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} \text{ and } V_S = \sqrt{\frac{\mu}{\rho}}$$

where V_P = *P* wave velocity in m/s, V_S = *S* wave velocity in m/s, K (bulk modulus) = $\rho (V_P^2 - \frac{4}{3} V_S^2)$, μ (shear modulus/rigidity modulus) = ρV_S^2 , ρ = density in gm/cc.

Elastic Wave Velocity as a Function of Geological Age and Depth

$$v = 1.47 (zT)^{\frac{1}{6}} \text{ in km/s.}$$

where z = depth in km, T = geological age in millions of years.

Time Average Equation to Estimate Rock Porosity

$$\frac{1}{v} = \frac{\varphi}{v_f} + \frac{(1-\varphi)}{v_m}$$

where v = *P* wave velocity for a rock, φ = fractional porosity, v_f = acoustic velocity in the pore fluid, and v_m = acoustic velocity in the rock.

RAY PATHS IN LAYERED MEDIA

Reflection and transmission of normally incident seismic rays

Seismic/acoustic impedance:

$$Z = \rho v$$

where ρ = density of the medium, v = *P* wave velocity.

Amplitude reflection coefficient:

$$R = \frac{A_1}{A_0} = \frac{Z_2 - Z_1}{Z_2 + Z_1}$$

Amplitude transmission coefficient:

$$T = \frac{A_2}{A_0} = \frac{2Z_1}{Z_2 + Z_1}$$

where A_1 and A_2 are the amplitudes of the reflected and refracted *P* waves, respectively, in terms of the amplitude A_0 of the incident wave. The energy of the wave is proportional to the square of its amplitude. The fraction E_r of the incident energy that is reflected is given by the square of R ; the fraction E_t that is transmitted is equal to $(1 - E_r)$.

Reflection and Refraction of the Obliquely Incident Rays

Four types of waves are generated: reflected and transmitted P waves, reflected and transmitted S waves. Snell's law to both the reflected and refracted P and S waves gives

$$\frac{\sin i_p}{\alpha_1} = \frac{\sin r_p}{\alpha_2} = \frac{\sin i_s}{\beta_1} = \frac{\sin r_s}{\beta_2} \Rightarrow \frac{\sin i_p}{\sin r_p} = \frac{\alpha_1}{\alpha_2}$$

Where:

i_p, i_s = angle between normal and ray paths of P and S wave in medium 1.

r_p, r_s = angle between normal and ray paths of P and S wave in medium 2.

α_1, β_1 = P and S wave velocities in medium 1.

α_2, β_2 = P and S wave velocities in medium 2.

Subcritical and Supercritical Reflections, and Critical Refraction

For a critical angle of incidence, the corresponding refracted ray makes an angle of refraction of 90° with the normal to the boundary. By applying Snell's law:

$$\frac{\sin i_c}{\sin 90^\circ} = \frac{\alpha_1}{\alpha_2} \Rightarrow \sin i_c = \frac{\alpha_1}{\alpha_2}$$

where i_c = critical angle of incidence, α_1 = P wave velocity in upper medium, α_2 = P wave velocity in lower medium. The critical reflection defines two domains corresponding to regions of subcritical and supercritical reflection (wide-angle reflection), respectively.

SEISMIC REFLECTION SURVEYING

Introduction and General Considerations

The most important problem in seismic reflection surveying is the translation of two-way travel time (TWTT) to depth. While travel times are measured, the one parameter that most affects the conversation of the depth is seismic velocity (two unknown depth and velocity).

Velocity is given by the following formula:

$$v_i = \frac{Z_i}{\tau_i}$$

where v_i = interval velocity, Z_i = thickness of the interval, τ_i = one way travel time. The interval velocity may be averaged over several depth intervals to yield an average velocity $\bar{v}$.

$$\bar{v} = \frac{Z_n}{T_n}$$

where Z_n = total thickness of the top n layers and T_n = total one-way travel time through the n layers.

Reflection at a Single Horizontal Reflector

Travel time of the direct wave:

$$t_{\text{direct}} = \frac{x}{v}$$

where v = P wave velocity, x = horizontal distance from shot point.

Travel time of the reflected wave:

$$t = \frac{2}{V} \sqrt{d^2 + \frac{x^2}{4}} = \frac{2d}{V} \sqrt{1 + \frac{x^2}{4d^2}} = t_0 \sqrt{1 + \frac{x^2}{4d^2}}$$

where d = depth to the reflecting bed, $t_0 (x = 0) = \frac{2d}{v}$ = echo time (vertical echo), $1 + \frac{x^2}{4d^2} =$ determines the curvature of t - x curve = normal move-out factor. Equation of the hyperbola:

$$\frac{t^2}{t_0^2} - \frac{x^2}{4d^2} = 1$$

Normal move-out:

$$\Delta t_n = t_x - t_0 = \frac{x^2}{2v^2 t_0}$$

Reflection at a Single Dipping Reflector

The travel time of the reflection from the inclined boundary:

$$t = \frac{1}{V} \sqrt{x^2 + 4xd\sin\theta + d^2}$$

where d = perpendicular distance to the inclined plane.

Minimum point of the travel time hyperbola:

$$x_m = -2d\sin\theta \text{ and } t_m = \frac{2d\cos\theta}{V}$$

Vertical and Horizontal Resolution (Fresnel Zone)

The vertical resolution of a seismic pulse is $\left(\frac{1}{4} \text{ to } \frac{1}{8}\right)$ of a wavelength. *Vertical resolution determines the thickness of the beds.*

Horizontal resolution is determined by the detector spacing and the Fresnel zone, where $w = (2z\lambda)^{\frac{1}{2}}$ for $z \gg \lambda$, z = reflection depth.

SEISMIC REFRACTION SURVEYING

Refraction at a Horizontal Interface

Travel time of the direct wave:

$$t_{\text{direct}} = \frac{x}{v_1}$$

Travel time of the doubly refracted ray:

$$t = \frac{x}{V_2} + \frac{2d}{V_1} \cos i_c$$

Intercept time:

$$t_i = \frac{2d}{V_1} \cos i_c = 2d \frac{\sqrt{V_2^2 - V_1^2}}{V_1 V_2}$$

Crossover distance (direct and doubly refracted rays):

$$x_{cr} = 2d \frac{\sqrt{V_2^2 - V_1^2}}{V_1 V_2}$$

where V_1 = velocity of the upper layer, V_2 = velocity of the lower layer, d = depth to the interface, i_c = critical angle.

Refraction at an Inclined Interface

The travel time for down-dip shooting:

$$t_d = \frac{x \cos \theta}{V_2} + \frac{d_A + d_B}{V_1} \cos i_c$$

Intercept time for the down-dip shot:

$$t_{id} = \frac{2d_A}{V_1} \cos i_c$$

The travel time for up-dip shooting:

$$t_u = \frac{x}{V_1} \sin(i_c - \theta) + t_{iu}$$

Intercept time for the up-dip shot:

$$t_{iu} = \frac{2d_B}{V_1} \cos i_c$$

True velocity of the lower layer:

$$\frac{1}{V_d} + \frac{1}{V_u} = \frac{2}{V_2} \cos \theta$$

If the refractor dip is small, $\cos \theta \approx 1$:

$$\frac{1}{V_2} \approx \frac{1}{2} \left(\frac{1}{V_d} + \frac{1}{V_u} \right)$$

where d_A = perpendicular distance from the shot point A, d_B = perpendicular distance from the shot point B, θ = refractor dip, and V_d and V_u are the apparent velocities.

EARTHQUAKE MAGNITUDE

Richter scale

$$M_S = \log_{10} \left(\frac{A_S}{T} \right) + 1.66 \log_{10} \Delta + 3.3$$

where A_S = amplitude of surface wave motion in microns; T = period of surface wave (around 20 seconds); and Δ = distance to earthquake epicentre (degrees).

Seismic moment:

$$M_0 = \mu D A$$

where μ = shear modulus, A = area of the rupture along the geologic fault where the earthquake occurred (in square metres), D = average slip (displacement offset between the two sides of the fault) on A (in metres). M_0 , measured in dyne-cm (1 dyne-cm = 10^{-7} newton-metre), is essentially the amount of energy released by the earthquake.

Moment magnitude:

$$M_w = \frac{2}{3} \log_{10} M_0 - 6$$

where suffix w stands for work (energy).

16.4.4 ELECTRIC AND ELECTROMAGNETIC

ELECTRICAL PRINCIPLES

Force of attraction or repulsion between two charged spheres (Coulomb's law):

$$F = k \frac{q_1 q_2}{r^2}$$

where q_1 and q_2 are the electric charges, r is their separation, and k is a constant. In SI system:

$$F = \frac{1}{4\pi\epsilon_0} \frac{q_1 q_2}{r^2}$$

where ϵ_0 = permittivity constant = $8.854 \times 10^{-12} \text{ C}^2 \text{N}^{-1} \text{m}^{-2}$.

Electric Field and Potential

$$E = \text{electric field} = \frac{q}{4\pi\epsilon_0 r^2} = -\frac{dU}{dr}$$

E in SI system (NC^{-1}).

$$U = \text{electric potential} = \frac{q}{4\pi\epsilon_0 r}$$

U in SI system (joules/coulomb) = volt, so E will be in V/m.

Ohm's Law

$$V = IR, R = \rho \frac{L}{A}$$

where R = resistance in ohm (Ω), V = potential difference, I = electric current, A = cross-sectional area of conductor, L = length of conductor, ρ = electrical resistivity or electrical specific resistance in $\Omega\text{-m}$. The inverse of resistance is called the conductance; unit is the reciprocal ohm (Ω^{-1}), also called a mho or siemens (S).

$$\sigma = \frac{1}{\rho} = \text{conductivity in } (\Omega^{-1} \text{m}^{-1}).$$

Ohm's law can be written as:

$$J = \sigma E$$

where J is in amp/m² (A m⁻²).

Electric displacement:

$$D = \epsilon_0 E + P = \epsilon_0 E + \chi_e E = (\epsilon_0 + \chi_e) E = \epsilon E$$

where P = electric polarization, χ_e = electric susceptibility, μ = electric permittivity.

Dielectric constant:

$$\eta = \frac{\epsilon}{\epsilon_0}$$

where $\epsilon = \epsilon_r \epsilon_0$, ϵ_r = relative permittivity. The dielectric constant is dimensionless.

PRINCIPLES OF RESISTIVITY METHOD

Electric potential in an infinite homogeneous and isotropic medium:

$$V = \frac{I\rho}{4\pi r}$$

For a semi-infinite medium:

$$V = \frac{I\rho}{2\pi r}$$

In direct current (DC) electrical prospecting:

$$\rho = \frac{2\pi}{\left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN}\right)} \frac{\Delta V}{I} = K \frac{\Delta V}{I}$$

where $K = \frac{2\pi}{\left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN}\right)}$ = geometric factor of the electrode arrangement.

Over a semi-infinite space of homogeneous and isotropic material, then the value of ρ computed earlier will be the true resistivity of that material. However, if the medium is inhomogeneous and/or anisotropic, then the resistivity computed earlier is called an apparent resistivity (ρ_a).

ELECTRODE CONFIGURATIONS

Wenner array:

$$\rho_a = 2\pi a \frac{\Delta V}{I}$$

where $AM = MN = NB = a$ = array spacing.

Schlumberger array:

$$\rho_a = \pi \frac{\left(\frac{AB}{2}\right)^2 - \left(\frac{MN}{2}\right)^2}{MN} \frac{\Delta V}{I}$$

Pole–dipole array:

$$\rho_a = 2\pi a n(n+1) \frac{V}{I}$$

where n is the separation parameter.

Three-electrode array:

$$\rho_a = 4\pi a \frac{V}{I}$$

Pole–pole array:

$$\rho_a = 2\pi a \frac{V}{I}$$

Dipole–dipole array:

The approximate geometric factors for different bipole–dipole configurations are as follows:

$$K_{\text{parallel}} = \frac{2\pi r^3}{Ll} \left[\frac{1}{3\cos^2\theta - 1} \right]$$

$$K_{\text{perpendicular}} = \frac{2\pi r^3}{3Ll} \left[\frac{1}{\sin\theta\cos\theta} \right]$$

$$K_{\text{radial}} = \frac{\pi r^3}{Ll\cos\theta}$$

$$K_{\text{azimuthal}} = \frac{2\pi r^3}{Ll\sin\theta}$$

$$K_{\text{equatorial}} = \frac{2\pi r^3}{Ll}$$

where $AB = L$, $MN = l$, r = dipole separation, θ = dipole angle.

Collinear dipole–dipole array:

$$\rho_a = \pi a n(n+1)(n+2) \frac{V}{I}$$

where a = dipole length.

EFFECT OF INHOMOGENEOUS GROUND

Distortion of Current Flow at a Plane Interface

Two homogeneous media of resistivities (ρ_1) and (ρ_2) separated by a plane boundary:

$$\frac{\tan\theta_2}{\tan\theta_1} = \frac{\rho_1}{\rho_2}$$

Where θ_1 and θ_2 are the angles to the normal.

Distortion of Potential at a Plane Interface

Reflection coefficient:

$$k = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}$$

RESISTIVITY SURVEYING

Electrical Sounding Curve

The curve of $\rho_a = f(a, \frac{AB}{2}, \text{ or } r)$ is called an electrical sounding curve.

Probing Depth

The greater the current electrode separation, the greater the amount of current that penetrates to a given depth. This relation is governed by the equation

$$\frac{I_s}{I_t} = \frac{2}{\pi} \tan^{-1} \left(\frac{2z}{AB} \right)$$

where I_s = current confined between depth (0) and (z), I_t = total current penetrating the ground, and AB = distance separating current electrodes.

Goelectric Parameters

Longitudinal unit conductance:

$$S_i = \frac{h_i}{\rho_i}$$

Transverse unit resistance:

$$T_i = h_i \rho_i$$

Longitudinal resistivity:

$$\rho_L = \frac{h_i}{S_i}$$

Transverse resistivity:

$$\rho_t = \frac{T_i}{h_i}$$

Anisotropy:

$$\lambda = \sqrt{\frac{\rho_t}{\rho_L}}$$

For an isotropic layer, $\rho_t = \rho_L$ and $\lambda = 1$.

Subscript i indicates the position of the layer in the section ($i = 1$ for the uppermost layer).

Total longitudinal unit conductance:

$$S = \sum_{i=1}^n \frac{h_i}{\rho_i} = \frac{h_1}{\rho_1} + \frac{h_2}{\rho_2} + \dots + \frac{h_n}{\rho_n}$$

Total transverse unit resistance:

$$T = \sum_{i=1}^n h_i \rho_i = h_1 \rho_1 + h_2 \rho_2 + \dots + h_n \rho_n$$

Average longitudinal resistivity:

$$\rho_L = \frac{H}{S} = \frac{\sum_{i=1}^n h_i}{\sum_{i=1}^n \frac{h_i}{\rho_i}}$$

Average transverse resistivity:

$$\rho_t = \frac{T}{H} = \frac{\sum_{i=1}^n h_i \rho_i}{\sum_{i=1}^n h_i}$$

Anisotropy:

$$\lambda = \sqrt{\frac{\rho_t}{\rho_L}} = \frac{\sqrt{TS}}{H}$$

The parameters T and S were named the ‘Dar Zarrouk’ parameters.

Types of Electrical Sounding Curves over Horizontally Stratified Media

The apparent resistivity curve for a three-layer structure (of resistivities ρ_1 , ρ_2 , and ρ_3 , and thicknesses h_1 , h_2 , and $h_3 = \infty$):

$\rho_1 < \rho_2 > \rho_3 \dots \dots \dots$ K-type section

$\rho_1 > \rho_2 < \rho_3 \dots \dots \dots$ H-type section

$\rho_1 < \rho_2 < \rho_3 \dots \dots \dots$ A-type section

$\rho_1 > \rho_2 > \rho_3 \dots \dots \dots$ Q-type section

The apparent resistivity curve for a four-layer structure (of resistivities ρ_1 , ρ_2 , ρ_3 , and ρ_4 , and thicknesses h_1 , h_2 , and $h_3 = \infty$):

$\rho_1 > \rho_2 < \rho_3 < \rho_4 \dots \dots \dots$ HA-type section

$\rho_1 > \rho_2 < \rho_3 > \rho_4 \dots \dots \dots$ HK-type section

$\rho_1 < \rho_2 < \rho_3 < \rho_4 \dots \dots \dots$ AA-type section

$\rho_1 < \rho_2 < \rho_3 > \rho_4 \dots \dots \dots$ AK-type section

$\rho_1 < \rho_2 > \rho_3 < \rho_4 \dots \dots \dots$ KH-type section

$\rho_1 < \rho_2 > \rho_3 > \rho_4 \dots \dots \dots$ KQ-type section

$\rho_1 > \rho_2 > \rho_3 < \rho_4 \dots \dots \dots$ QH-type section

$\rho_1 > \rho_2 > \rho_3 > \rho_4 \dots \dots \dots$ QQ-type section

Induced Polarization (IP)

Chargeability:

$$M = \frac{1}{V_0} \int_{t_1}^{t_2} V(t) dt$$

The ratio $\frac{V_t}{V_0}$ is expressed as mV/V, and M has the dimensions of time and is expressed in seconds or milliseconds. V_0 = steady-state voltage, $V(t)$ = decaying residual voltage.

Frequency effect:

$$FE = \frac{\rho_f - \rho_F}{\rho_F}$$

The ratio FE is often multiplied by 100 to express it as a percentage (PFE). Where ρ_f = apparent resistivity, measured at $f \approx 0.05$ –5 Hz, and ρ_F = apparent resistivity, measured at $F \approx 1$ –10 Hz.

Metal factor:

$$MF = A (\sigma_F - \sigma_f) = A \left(\frac{1}{\rho_F} - \frac{1}{\rho_f} \right) = A \left(\frac{\rho_f - \rho_F}{\rho_f \rho_F} \right)$$

The constant A is equal to $2\pi \times 10^5$; the units of MF are those of conductivity (i.e. $\Omega^{-1}\text{m}^{-1}$ or S m^{-1}).

Relation between time domain (TD) and frequency domain (FD) IP measurement:

$$MF = \frac{\text{PFE}}{100 + \text{PFE}}$$

Self-induced polarization (SIP):

$$\text{Complex impedance } Z(\omega) = \frac{U(\omega)}{I(\omega)} K$$

where K = geometric factor of the array, $U(\omega)$ = voltage amplitude, $I(\omega)$ = current amplitude.

The Cole–Cole model for complex resistivity:

$$Z(\omega) = \rho_0 \left[1 - M \left(1 - \frac{1}{(1 + i\omega\tau)^c} \right) \right] = |Z| e^{i\varphi}$$

where ρ_0 is the DC resistivity, M is the chargeability, and τ is the relaxation time.

The critical frequency at which the maximum phase shift is measured is indicative of (τ):

$$F_c = \frac{1}{2\pi(1-M)^{\frac{1}{2c}}}$$

where $\tau = \frac{1}{2\pi}$, $c = 0.5$, and $M = 0$ –1.

ELECTROMAGNETIC (EM) INDUCTION

Skin Depth

$$\delta = \sqrt{\frac{2}{\omega \mu \sigma}}$$

where $\omega = \frac{2\pi}{T}$, T = time period, μ = magnetic permeability, σ = conductivity.

In-Phase Component and Quadrature Component (Transmitter–Receiver System) Used in Horizontal Loop EM Method

The voltage ratio:

$$\frac{V_s}{V_p} = \frac{M_{cr} I_c}{M_{tr} I_t} = -\frac{M_{tc} M_{cr}}{M_{tr} L} \left(\frac{\beta^2 + i\beta}{1 + \beta^2} \right) = \text{response of the measuring system}$$

where $\beta = \frac{\omega L}{R}$ = response parameter of the conductor; V_s = secondary voltage; V_p = primary voltage; M_{cr} = mutual inductance between conductor and receiver; M_{tr} = mutual inductance between transmitter and receiver; I_c = current system in conductor; I_t = current system in transmitter; M_{tc} = mutual inductance between transmitter and conductor; L = self-inductance of the conductor; and $\beta = \frac{\omega L}{R}$ is the response parameter of the conductor, with R = resistance of the conductor.

The voltage ratio:

$$\frac{V_s}{V_p} = P + iQ$$

where real part P has the same phase as the primary signal and is called the in-phase component of the response. The imaginary part Q is 90° out of phase with the primary signal and is called the quadrature component of the response.

Ground Conductivity Meters (Measurements at Low Induction Number, Coil Space Smaller Than the Skin Depth)

The in-phase component of the signal becomes very small, and the quadrature component is directly related to the conductivity of the ground:

$$\sigma_a = \left(\frac{4}{\mu_0 \omega s^2} \right) \left(\frac{H_s}{H_p} \right)_q$$

where subscript q denotes the quadrature phase, intercoil separation s , H_s the secondary magnetic field, H_p is the primary magnetic field, and σ_a is the apparent conductivity of the subsurface.

GROUND-PENETRATING RADAR (GPR)

Contrasts in dielectric properties of rocks set the basic working principle of the GPR method. Material properties relevant for electromagnetic wave propagation are the electric permittivity ϵ and the specific electric conductivity A .

Propagation Speed of EM Wave

The propagation speed of electromagnetic waves in the subsurface:

$$v = \frac{c}{\sqrt{\epsilon}}$$

where c = speed of light in vacuum = 3×10^8 m/s. The propagation speed of electromagnetic waves is used for time/depth conversion of GPR sections.

Electric permittivity and conductivity:

Dielectric constant, $\eta = \frac{\epsilon}{\epsilon_0}$ where $\epsilon = \epsilon_r \epsilon_0$ = electric permittivity, ϵ_r = relative permittivity.

To be consistent with the usual notation in electrodynamics, we introduce a quantity called complex permittivity ϵ^* :

$$\epsilon^* = \epsilon' - i\epsilon''$$

Normalization with the vacuum permittivity ϵ_0 yields the dimensionless relative complex permittivity:

$$\epsilon_r^* = \frac{\epsilon^*}{\epsilon_0} = \epsilon_r' - i\epsilon_r''$$

The loss tangent is the ratio of conduction to displacement currents and defined as:

$$\tan \delta = \frac{\epsilon_r'}{\epsilon_r''} = \frac{\sigma}{2\pi f \epsilon_0 \epsilon_r}, \quad \epsilon_r'' = \frac{\sigma}{\omega \epsilon_0}$$

EM wave propagation utilized in GPR takes place for $\tan \delta < 1$, while for $\tan \delta > 1$, diffusive processes govern (the case of low-frequency EM).

The wavelength λ is calculated from the well-known relation $v = \lambda f$ for low-loss materials as:

$$\lambda = \frac{c_0}{f \sqrt{\epsilon}} \quad \text{where } c_0 = 3 \times 10^8 \text{ m/s.}$$

Reflection and Refraction of Plane EM Waves

According to Snell's law:

$$\frac{\sin \theta_i}{\sin \theta_r} = n_{21} = \sqrt{\frac{\epsilon_2}{\epsilon_1}} = \frac{v_1}{v_2}$$

where n_{21} = index of refraction, θ_i = angle of incidence, θ_r = angle of refraction.

Coefficient of reflection (reflectivity) (r). Reflected part of incident unit amplitude (r) is given by:

$$r = \frac{\sqrt{\epsilon_1} - \sqrt{\epsilon_2}}{\sqrt{\epsilon_1} + \sqrt{\epsilon_2}} = \frac{v_2 - v_1}{v_2 + v_1}$$

Coefficient of transmission (t). Refracted (transmitted) part (t) is given by:

$$t = \frac{2v_2}{v_2 + v_1} = \frac{2\sqrt{\epsilon_1}}{\sqrt{\epsilon_1} + \sqrt{\epsilon_2}}$$

where ε_1 and ε_2 are the relative permittivities (dielectric constants) of the two media separated by the interface, and v_1 and v_2 the radar velocities within them.

Equations for Signal Travel Times

$$\text{Direct air wave: } t_a = \frac{x}{c_0}$$

$$\text{Direct ground wave: } t_g = \frac{x}{v}$$

$$\text{Reflected wave: } t_r = \frac{1}{v} \sqrt{x^2 + 4h^2}$$

$$\text{Critically refracted wave (lateral wave): } t_l = \frac{x}{c_0} + \sqrt{\frac{1}{v^2} - \frac{1}{c_0^2}}$$

where c_0 = speed of EM wave = speed of light; v = velocity in the uppermost stratum; x = distance; h = depth to the layer boundary. The travel times are calculated in nanosecond (ns) if the velocities are in m/ns and the distances are in m.

Amplitudes/Attenuation

$$\frac{E_r}{E_l} = \frac{r}{2h} e^{-2\alpha h}$$

where E_r = amplitude of reflected wave, E_l = amplitude at unit depth, r = reflectivity, α = attenuation coefficient, which is dependent upon the electric conductivity σ , magnetic permeability μ , and dielectric permittivity ε properties of the geological medium as well as the frequency of the radar signal. Smaller value of α for air, ice, granite provides an ideal condition for deploying GPR.

In the case of diffractor:

$$\frac{P_r}{P_t} = q_s \frac{G^2 \lambda^2}{(4\pi)^3 h^4} e^{-4\alpha h}$$

where P_t = transmitted power (in watt), P_r = received power (in watt), q_s = scatter cross section (m^2), G = antenna gain, λ = wavelength at the centre frequency (in m), h = distance between the antenna and diffractor (m), α = attenuation coefficient (in m^{-1}).

Estimation of Exploration Depth

The maximum exploration depth:

$$h_{\max} \leq 0.018 \rho \sqrt{\varepsilon}$$

16.4.5 MAGNETOTELLURIC (NATURAL SOURCE LOW-FREQUENCY EM)

PRINCIPLES OF ELECTROMAGNETIC (EM) SURVEYING

Maxwell equations:

$$\nabla \times \mathbf{E} = - \frac{\partial \mathbf{B}}{\partial t}, \text{ Faraday's law, } E \text{ in V/m and } B \text{ in tesla}$$

$$\nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t}, \text{ Ampere's law, } H \text{ in A/m, } J \text{ in A/m}^2, \text{ and } D \text{ in C/m}^2$$

$$\nabla \cdot \mathbf{B} = 0, \text{ Gauss's law for magnetism}$$

$$\nabla \cdot \mathbf{D} = \rho, \text{ Gauss's law for electrostatic}$$

$$\nabla \cdot \mathbf{J} + \frac{\partial \rho}{\partial t} = 0, \text{ equation of continuity}$$

In the presence of a linear isotropic medium, the material equations:

$$\mathbf{B} = \mu \mathbf{H}$$

$$\mathbf{D} = \epsilon \mathbf{E}$$

$$\mathbf{J} = \sigma \mathbf{E}$$

where $\mu = \mu_0 \mu_r$ = magnetic permeability, and $\epsilon = \epsilon_0 \epsilon_r$ = dielectric permittivity, with $\mu_0 = 4\pi \times 10^{-7}$ H/m-magnetic permeability of free space and $\epsilon_0 = 8.85 \times 10^{-12}$ F/m-dielectric permittivity of free space. Variations in μ_r and ϵ_r are assumed negligible compared to variations in the bulk conductivity σ of rocks.

Helmholtz wave equation:

$$(\nabla^2 - K^2) \mathbf{B} = 0 \text{ for B field}$$

$$(\nabla^2 - K^2) \mathbf{E} = 0 \text{ for E field}$$

$$\text{For example: } \frac{\partial^2 H_x}{\partial z^2} = K^2 H_x \text{ and } \frac{\partial^2 E_x}{\partial z^2} = K^2 E_x \text{ etc.}$$

where $K = \sqrt{i\omega\mu\sigma}$ = propagation constant, and $i = \sqrt{-1}$, the imaginary number. We use H instead of B for describing the magnetic field, since E and H are physically analogous fields, which is more apparent.

Skin Depth

The inverse of the real part of K is known as the skin depth or penetration depth (δ).

$$\delta = \frac{1}{\text{Re}K} = \sqrt{\frac{2}{\omega\mu\sigma}} \Rightarrow \delta(\text{T}) = 504 \sqrt{T\rho_a} \text{ in (m)}$$

where $\rho_a = \frac{1}{\sigma_a}$, $T = \frac{2\pi}{\omega}$, period in (s), $T = \frac{1}{f}$, where f is frequency. EM fields (magnetic) are attenuated to a value of e^{-1} ($\sim 37\%$) of their surface amplitude at a depth δ (T) outside the conductor.

Apparent Resistivity and Phase of EM Wave (Natural Source MT)

Magnetotelluric (MT) impedance. The magnetotelluric impedance (Z) (in m/s) for a plane wave is the ratio of the electric and magnetic fields measured and oriented orthogonally to each other at the surface:

$$Z = \frac{E}{H}$$

The resistivity of the half-space, in units of $\Omega\text{-m}$:

$$\rho = \frac{1}{\omega\mu} |Z|^2 = \frac{T}{2\pi\mu} |Z|^2 = 0.2T \frac{|E_x|^2}{|H_y|^2}$$

When the Earth resistivity is nonuniform, the right-hand sides of the preceding equation provide apparent resistivities (instead of true resistivity), ρ_a , which are frequency (period)–dependent, as is the case with 1-D, 2-D, or 3-D situations.

In the frequency domain, $Z(\omega)$ is complex and has an associated phase (ϕ):

$$\phi = \tan^{-1} \frac{\text{Im} \left| \frac{E_x}{H_y} \right|}{\text{Re} \left| \frac{E_x}{H_y} \right|}$$

Anisotropic medium:

$$\begin{pmatrix} j_x \\ j_y \\ j_z \end{pmatrix} = \begin{pmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz} \end{pmatrix} \begin{pmatrix} E_x \\ E_y \\ E_z \end{pmatrix}$$

That is, Ohm's law is expressed in 3-D tensor form.

Impedance tensor:

$$Z = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix}$$

MT Response Functions

1. Impedance Tensor

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}$$

For 1-D Earth:

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} 0 & Z_{xy} \\ -Z_{yx} & 0 \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}$$

For 2-D Earth:

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}$$

For 3-D Earth:

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}$$

The tensor (Z) can be rotated to any other coordinate system (Z_m) by an angle (θ) with the rotation matrix (R).

$$Z_m = RZR^T \text{ where } R = \begin{pmatrix} \cos\theta & \sin\theta \\ -\sin\theta & \cos\theta \end{pmatrix}$$

The phase of the impedance element means the phase shift between the electric and magnetic field components:

$$Z = \left| \frac{E_i}{H_j} \right| e^{i\varphi}$$

$$\varphi = \arctan \frac{\text{Im}Z_{ij}}{\text{Re}Z_{ij}}$$

In a homogeneous Earth, the impedance phase $\varphi = 45^\circ$.

2. Directionality Parameter

Strike. The direction in which the conductivity of a 2-D structure does not vary indicates the strike direction (principal conductivity axis). The angle between the principal conductivity axis and the x axis is called the angle of strike.

The angle of strike:

$$\theta_0 = \frac{1}{4} \arctan \frac{(Z_{xx} - Z_{yy})(Z_{xy} + Z_{yx})^* + (Z_{xx} - Z_{yy})^*(Z_{xy} + Z_{yx})}{(Z_{xx} - Z_{yy})^2 - (Z_{xy} - Z_{yx})^2}$$

where $*$ denotes the complex conjugate.

3. Dimensionality Indicators

Skew. Skew is defined as the ratio of the magnitude of the second invariant to that of the third invariant.

$$S = \frac{|Z_{xx} + Z_{yy}|}{|Z_{xy} - Z_{yx}|}$$

where Z_{xx} , Z_{xy} , Z_{yx} , and Z_{yy} are the impedance tensor elements. *Skew* is the measure of the EM coupling between the measured electric and magnetic field variations in the same direction.

Tipper:

$$H_z = AH_x + BH_y$$

where H_z = vertical magnetic component, and A and B are tipper coefficients.

The magnitude of the tipper is given by

$$T = \sqrt{A^2 + B^2}$$

The angle which maximizes the coherency between the horizontal and vertical magnetic fields is given by tipper angle:

$$\varphi = T^{-2} \left(\text{Re}^2 A + \text{Re}^2 B \right) \arctan(\text{Re} B + \text{Re} A) + (\text{Im}^2 A + \text{Im}^2 B) \arctan(\text{Im} B + \text{Im} A)$$

Tipper skew is given by:

$$\gamma = 2T^{-2}(\text{ReAImB} - \text{ImAReB})$$

For 2-D structures, γ is zero. All three quantities T , φ , and γ are independent of axis rotation and provide some information about the subsurface structure.

4. Induction Arrows (Tipper Vector)

$$\mathbf{H}_z(\omega) = \left(\mathbf{T}_x(\omega) \mathbf{T}_y(\omega) \right) \begin{pmatrix} \frac{\mathbf{B}_x}{\mu_0} \\ \frac{\mathbf{B}_y}{\mu_0} \end{pmatrix}$$

Induction arrows can be used to infer the presence or absence of lateral variations in conductivity. Induction arrows are vector representations of the complex (i.e. containing real and imaginary parts) ratios of vertical to horizontal magnetic field components.

Static Shift

For long periods and a large skin depth (greater than the dimensions of the anomaly), the body will act as a local near-surface inhomogeneity, which gives a noninductive (galvanic) response. This produces a frequency-independent distortion called static shift.

3-D Galvanic Distortion and Decomposition of MT Impedance Tensor

The electromagnetic signal becomes weak when the electromagnetic skin depth becomes greater than the dimensions of the anomaly, but it continues to have a noninductive response, termed as galvanic. The decomposition technique is used to determine the electromagnetic strike by factorizing the impedance tensor problem in terms of a rotation matrix β and a distortion tensor C .

$$\mathbf{Z} = \beta \mathbf{C} \mathbf{Z} \beta^T, \text{ where } \mathbf{C} = \mathbf{g} \mathbf{T} \mathbf{S} \mathbf{A}$$

$$\beta = \begin{pmatrix} \cos\alpha & \sin\alpha \\ -\sin\alpha & \cos\alpha \end{pmatrix} \text{ and } \beta^T = \begin{pmatrix} \cos\alpha & -\sin\alpha \\ \sin\alpha & \cos\alpha \end{pmatrix}$$

where g = scalar, T = tensor suboperator twist, S = tensor suboperator shear, A = local anisotropy.

Rotation-Invariant Magnetotelluric Impedance Tensors

Rotation-invariant MT parameters are those whose amplitudes remain invariant as the MT tensors are rotated mathematically. It is an elegant method that did not require presumptions about the underlying dimensionality.

$$\mathbf{Z}' = \beta \mathbf{Z} \beta^T$$

The invariant is equal to

$$Z_{xx} Z_{yy} - Z_{xy} Z_{yx}$$

Apparent resistivity:

$$\rho_{\text{invariant}} = \frac{1}{\omega \mu_0} (Z_{xx} Z_{yy} - Z_{xy} Z_{yx})$$

Magnetotelluric Phase Tensors

MT phase tensors did not require presumptions about the underlying dimensionality. That means, the MT phase tensor is applicable where both the regional conductivity distribution and the conductivity heterogeneity close to the surface are 3-D.

Phase tensor:

$$\varphi = X^{-1}Y$$

where X^{-1} is the inverse of the real part and Y is the imaginary part of the impedance tensor (Z), respectively. The phase tensor is fully defined with the angle (α) and the three coordinate invariants φ_{max} (maximum phase tensor), φ_{min} (minimum phase tensor), and β (skew angle).

$$\varphi = R^T (\alpha - \beta) \begin{pmatrix} \varphi_{\text{max}} & 0 \\ 0 & \varphi_{\text{min}} \end{pmatrix} R (\alpha + \beta)$$

where $R(\alpha + \beta)$ = rotation matrix, $R^T(\theta) = R(-\theta)$, $(\alpha - \beta)$ = orientation of major axis of the tensor.

$$R(\alpha + \beta) = \begin{pmatrix} \cos(\alpha + \beta) & \sin(\alpha + \beta) \\ -\sin(\alpha + \beta) & \cos(\alpha + \beta) \end{pmatrix}$$

16.4.6 MARINE GEOPHYSICS

PROPAGATION OF SEISMIC WAVES THROUGH WATER

The relation between frequency (f) and wavelength (λ) is given by

$$\lambda = \frac{v}{f}$$

where λ is in m, f is in Hz, and v (velocity) is in m/s.

The P waves move away from the source and reflectors by alternatingly compressing and dilating the elastic media (water or sediments) along the travel path.

$$V_p = \sqrt{\frac{k + \frac{4}{3}\mu}{\rho}}$$

where V_p is the compressional wave velocity, k is bulk modulus, μ is the rigidity modulus or shear modulus, and ρ is the density of the media through which the wave is traveling.

The shear or S wave velocity:

$$V_s = \sqrt{\frac{\mu}{\rho}}$$

Due to the lack of elasticity of water, however, shear waves do not propagate through the water column.

16.5 PHYSICAL PROPERTIES OF EARTH MATERIALS

The degree of resolution from a particular geophysical method is contingent upon the variations of the subsurface properties being investigated. Table 16.3 is a simplified summary of the properties of common Earth materials. Variations of these properties along with the sensitivity of the measurement system determine the applicability of a geophysical method for a particular problem.

For reference, $\mu_0 = 4\pi \times 10^{-7}$ H/m; $\epsilon_0 = 8.8 \times 10^{-12}$ F/m

As seen in Table 16.3, densities of Earth materials are similar to each other. The highest densities are associated with ore bodies, whereas the lowest densities are associated with air (a subsurface cavity).

Magnetic susceptibilities of rocks are generally very weak ($\sim 10^{-4}$) and show large variations (from virtually no susceptibility to values 3–4 orders of magnitude higher than shown in Table 16.3). Rock susceptibility is almost entirely determined by its concentration of mineral magnetite and also the grain sizes of these magnetite minerals. With the highest magnetic contents with the smallest grain sizes, basalt has the highest magnetization among Earth materials. Lastly, the low magnetic susceptibilities of rocks ($\sim 10^{-4}$) allow the geophysicists who are modelling the Earth's electrical structure to take $\mu = \mu_0$ in most of the cases.

In terms of resistivities, we observe orders of magnitude differences among Earth materials. As metal-bearing deposits, oxides and sulphides have the lowest resistivities, but they are found only in localized areas. Among the commonly found constituents of the Earth subsurface, water and shale stand out with the lowest resistivities. Water conductivity can vary greatly depending on the dissolved salt content. Due to the high electrical conductivity of water, electric conduction in porous or fractured rocks is dominantly due to electrolytes inside the pores and fractures. As a result of

TABLE 16.3

Average Physical Properties of Some Materials Encountered in Geophysical Applications

			Magnetic Susceptibility		Dielectric	Seismic
	Material [#]	Density (gm/cc)	$\left(\frac{\mu}{\mu_0} - 1 \right)$	Log Resistivity (Ohm-m)	Constant $\left(\frac{\epsilon}{\epsilon_0} \right)$	Velocity (km/sec)
Various	Air	0.001	0	15	1	0.3
	Water	1.0	-7×10^{-10}	0–2	80	1.4–1.5
	Ice	0.9	-7×10^{-10}	6	3–4	3.4
	Oil	0.6–0.9	2×10^{-5}	14	2	1.3
	Salt	2.2	-1×10^{-6}	15	6	4.5–5
Unconsolidated sediments	Soil	1.5	7×10^{-4} *	3	4	0.1–0.2
	Clastics	1.9	5×10^{-4} *	3–4	4	1–2
	Sand	1.6	5×10^{-4} *	4	4	3
Metal ores	Oxides	3.8–9.1	3×10^{-3}	(–1) to 2	10–25	5.8
	Sulphides	3.8–8.1	3×10^{-3}	(–6) to (–3)	8–31	5.5
Sedimentary rocks	Sandstone	2.2	4×10^{-4} *	2–3	5	2–6
	Shale	2.7	6×10^{-4} *	0–1	6–8	2.3
	Limestone	2.1	3×10^{-4} *	2–3	8–9	3–6
Igneous rocks	Granites	2.6	2×10^{-3} *	4–6	5	5–6
	Basalt	3.0	7×10^{-2} *	7	12	5–6
Metamorphics	All	2.6–2.7	5×10^{-3} *	3–5	8–10	5.5–6

[#] For dry samples except water.

* May show orders of magnitude variations.

Source: Data compiled from Clark (1966); Telford et al. (1976); Parasnis (1986); Zhdanov and Keller (1994).

this, the bulk resistivity of rocks is expressed as a function of its water saturation, which is given by Archie's law (Zhdanov and Keller, 1994).

16.6 COMPARISON OF THE ELECTRICAL METHODS

In all electrical methods, various forms of the Maxwell equations were applied according to the selection of the source frequency. Table 16.4 shows a comparative summary of the electric methods used in geophysical applications.

The main characteristic of the electrical methods is controlled by the frequency ranges used during investigations. The source current is ideally constant in DC methods. In practicality, very slowly varying currents associated with very low frequencies are used to remove the effect of charge accumulation between electrodes and the ground. In DC methods, most of the electric energy is taken by the resistive layers, so these methods are more responsive to resistive anomalies.

The diffusion form of the EM fields (including EM, TEM, and MT) occurs at low frequencies (less than a MHz), which is being used where diffusive behaviour (quasistationary form) of the fields dominates, which in turn allows to make the dielectric properties of the Earth associated with the displacement currents least significant. Unlike the DC method, EM/TEM methods work best for conductive anomalies, as these methods use the principle of electromagnetic induction of conductors. Using the complementary behaviour of the DC and EM methods, the former being responsive to resistive layers, the latter being responsive to conductive layers, a combined interpretation procedure of both methods of DC and EM to improve the resolution of electrical inversions is described by Raiche et al. (1985).

The electromagnetic fields act as true waves in the Earth, and the electromagnetic fields act as true waves in the Earth at high frequencies (above MHz), for which the radiation forms of Maxwell's equations are applied. In this state, the dielectric properties of the medium dominate over the conductive properties in the behaviour of the EM field. Earth's conductivity only comes into the formulas in order to study the attenuation of the EM waves within the medium. In GPR, one uses the reflective properties of EM waves across material boundaries, which is a function of the contrast in dielectric (and magnetic) properties. The GPR method finds applications for both high and low dielectric anomalies, and the fast attenuation of waves occurs at high frequencies, which is why the GPR method is limited to very shallow applications.

16.7 MULTIPLE METHODS AND SURVEY DESIGN

As mentioned in previous chapters, more than one geophysical method can be used to help understand the geological context of the geophysical survey results. That is why an electromagnetic survey helps identify a conductor, but graphite is known to occur in the area. A small gravity survey

TABLE 16.4

A Comparison of Electrical Methods

Method	Frequency Range	Responsive Units	Field Behaviour	Effective Parameter	Practical Depths	Resolution
DC	N/A	Resistive	Stationary	σ	<300 m	0.5–10 m
IP	N/A	Both	Stationary	σ	<300 m	0.5–10 m
EM	1 Hz–100 KHz	Conductive	Quasistationary	σ	<5 km	0.5–10 m
TEM	10 mHz–100 KHz	Conductive	Quasistationary	σ	<10 km	0.5–10 m
MT	1 mHz–100 KHz	Both	Quasistationary	σ	0.1–100 km	1–10 m
GPR	10 MHz–1,000 MHz	Both	Radiation	ϵ, σ	<20 m	0.1–10 m

can be carried out to see if the EM conductor is associated with higher-density material. It is documented that in the absence of the gravity anomaly, the conductor is not likely to arise from metallic sulphides.

The gravity method (including airborne gravity) is being used extensively in the search for iron ore. It is not difficult to identify surface outcrop of iron formations with a magnetometer. There are magnetic rocks with variations of magnetic property associated with the magnetic response variability due to exposed or subcropping ore. If the strong response masks any of the responses of ore at depth, then it is not possible to determine how extensive the ore is. If the rocks are highly deformed and folded, any realistic modelling of the magnetic data would be next to impossible. The gravity method is useful for distinguishing the wheat from the chaff; it highlights the areas having significant subsurface accumulations.

Some targets, such as gold in quartz, may not have a geophysical property to help in exploration. Even if it becomes possible to find an indirect property, such as a property of the host rocks, it will help in exploration. For example, the quartz rocks hosting gold may have very high resistivity values. In this case, a resistivity/IP survey may help in mapping the distribution of high-resistivity quartz, which would help concentrate exploration activity on suitable host rocks.

As time passes, the fashion in mineral exploration gets changed, and targets which have been dismissed in the past may become attractive in the future. An example is graphite, which for many years has been considered a misleading pest in mineral exploration. The recent discovery of high-grade graphite deposits has turned graphite from hateful villain to desirable target.

16.8 GENERAL ASPECTS OF GEOPHYSICAL FIELDWORK

Until the early 1990s, most geophysical surveying was for resources—oil and gas, coal, base metal, and precious metals. Recently, environmental applications have become popular, and engineering applications have increased. In general, these last two applications mean that the targets are much shallower than for resources (they are sometimes categorized as ‘near-surface’ geophysics). Archaeological applications are also ‘near surface’, as is the growing application to forensic investigations.

16.8.1 CHOICE OF METHOD

The type of requirement will primarily determine which method, or methods, is optimum. Some methods are better suited to certain requirements than others. Seismic reflection is by far the most used method on petroleum targets because of its superior penetration and resolution capabilities.

It is important to use the methods that respond to the particular physical contrasts involved, for example, massive sulphide deposits do not respond as well to IP as they do to EM. On the other hand, IP is more responsive than EM on disseminated sulphide targets. A method will not be useful if there is no contrast in the particular property on which it operates, for example, if there is no detectable density change in the area, gravity should not be used.

Each method has particular capabilities and advantages and some limitations and disadvantages. Factors such as resolving power, speed, and cost are some of the considerations in choosing which method. One method may not give as much information as another, but the sacrifice in detail may be deemed acceptable because it is a lot faster or simpler to use. EM is a much faster way to measure ground conductivity than the DC electrical methods because the latter requires contact to the ground whereas EM methods do not. On the other hand, DC resistivity is less affected by high-frequency noise. All noncontact methods have the advantage of faster coverage. Noncontact methods should be preferred in the presence of contaminated ground. Certain methods are better for layered targets, such as underground water, and in this case, some methods have been developed especially to detect water.

16.8.2 INSTRUMENTATION

The ‘outdoors’ is very different to a laboratory as a place to try to make accurate measurements. As for operator comfort, it can be too hot, too cold, raining, blowing a gale, and dusty. The equipment also has to contend with these factors. The instrument case and connectors need to be sealed from moisture and dust, and the electronics able to withstand a wide range of temperatures. LCD screens, for example, can fail if they are too hot, and they are also affected by direct sunlight. In addition, the field areas are often remote from power, repair facilities, and sometimes the only means of transport is carrying by the operator.

The term ‘equipment’ refers to the entire system of instruments and their ancillary items, such as battery packs and cables, and the term ‘instrument(s)’ refers to the main measurement unit(s). Often, with newer instruments, the recording and display part of an instrument is a separate laptop computer. Geophysical field instruments generally should have the attributes as ruggedness (e.g. gravimeters are almost too delicate for field use and need special care to transport) and portability (in general, instruments are made to be as light as possible and usually have a carrying harness or handles).

16.9 INSTRUMENT AT THE FIELD SITE

16.9.1 SPECIFIC OPERATIONAL PRECAUTIONS

Some precautions are specific to the method. The ability of the particular instrument to detect unwanted responses is, to some extent, dependent on its level of sensitivity and needs to be judged individually. With the magnetic method, care should be taken not to have any magnetizable metal close to the sensor. Metal items of clothing, such as belt buckles, and even clasps in baseball caps, can be detected with the more sensitive, new instruments and should be removed if suspected of causing effects. ‘Cultural’ objects such as sheds, fences, etc. should be avoided. Vehicles should be moved out of effect.

In the case of gravity surveying, only very large masses, such as nearby terrain, can produce additional responses, but vibration, such as from wind or traffic, should be avoided. In windy conditions, the meter should be shielded in some way to deflect the wind. When surveying with gravimeters in cities, readings are often made at night, when the traffic has ceased or is minimal. For ground radar, objects on the surface near the antennae can produce reflections that can be confused with targets in the subsurface. With electrical methods, the main precaution is to avoid electrocution of personnel. Electrode arrays should not be set up near long wire fences, especially subparallel to them. This also applies to parallel sets of cables, especially if one line is carrying current and the other is measuring voltage. In this case, unwanted voltage can be induced into the voltage line by the process of ‘coupling’. Surface water will produce anomalous electrical readings.

For the electromagnetic method, any surface object that can be ‘electromagnetized’ (or ‘induced’) should be avoided. A common such source are coils of cable near the receivers. In the case of frequency domain EM (as distinct from time domain EM), the exact spacings between separate transmitter and receiver are critical, as they can affect the measurement by large percentages. For seismic surveys, if explosives are used as a source, fully trained and licensed operators should manage them. In fact, explosives should be avoided in favour of alternative sources where possible.

16.10 SELECTION OF SURVEY PARAMETERS

Readings should preferably be acquired on a regular grid, although this is not always necessary or expedient. Before deciding on the optimum parameters, such as line spacing and direction, reading

spacing, etc., it may be appropriate in the first instance to conduct a ‘reconnaissance’ of the area to be surveyed with a limited number of readings. This ‘trial’ survey can allow a decision to be made, firstly, as to whether it will be worthwhile at all to conduct a more ‘detailed’ survey, and then what parameters are best. A well-defined grid can, if needed, be incorporated into a wider coordinate system and allow for the matching and overlaying of the data with other information.

If satellite navigation instruments such as ‘GPS’ receivers are used to define positions, the global coordinates they read can be used. The use of such receivers is particularly expedient for reconnaissance surveying or when not using a grid. In the case of multiple electrode or geophone arrays, the coordinates of the individual positions can be noted, particularly if the spacings are large, or just the centre position noted together with the spacing of the electrodes. For large EM loops, the corners can be assigned their appropriate coordinate values.

16.10.1 TRAVERSE OR LINE SPACING

The line spacing should be such as to have at least three lines crossing the anomaly (as distinct from the target, which may have a smaller spatial extent), and for further expediency, it is best that they are parallel and across the widest part of the anomaly. In the ideal case of a sphere, for example, the spacing must equal or be less than the diameter of the sphere.

16.10.2 STATION SPACING

A similar compromise exists between reading too close together, which takes up time and may not always exhibit new information, and the need to get optimum coverage of the anomaly. In this case, there should be sufficient readings on the anomaly to resolve its full shape. If the spacing is too great, the shape will be too smooth. This is a case of insufficient sampling of data, and the result is termed ‘aliasing’ of data.

16.10.3 ORIENTATION OF LINES

For maximum efficiency, lines should be oriented so as to gain the greatest amount of change over the anomaly. If the anomaly is very elongated, lines perpendicular to the axis of elongation are optimum. Lines parallel to the axis of elongation are in the worst direction as they will read essentially the same values along their length.

16.11 FIELD READINGS

16.11.1 SENSITIVITY, PRECISION, AND ACCURACY

Instrument specifications may quote values for one or more of these parameters. It is important to understand the differences.

1. *Sensitivity*. The sensitivity of an instrument is its ability to detect changes in measurement values, for example, gravimeters are sensitive to 1 part in 10^{-9} of the Earth’s average gravity value. The value of the very smallest change detectable is the ‘resolution’ of the instrument.
2. *Precision*. The precision of an instrument is a measure of its ability to give close ‘repeatability’ of a number of readings, for example, ‘drift’ due to temperature, or whatever.
3. *Accuracy*. The accuracy is the closeness of a reading to its true value. Accuracy can be improved by repeated measurements.

16.11.2 ERRORS

The result of any measurement can never be absolutely correct for a number of, mainly practical, reasons, and hence, there is always some error. The 'reliability' to be placed on a measurement can be expressed in terms of this error.

Types of Error

1. *Accountable error*. Systematic error (instrument wrongly calibrated), parallax error (reading incorrectly).
2. *Unaccountable error*. They are 'random' and may show up as a nonsensical variation, for example, radiometric method, because it measures random signals from the effects of cosmic radiation, requires multiple readings to be taken at each station, as the final value at any place is necessarily a statistical average.

In general, results can be improved and 'anomalies' better defined by acquiring more readings in a given distance or space. In this procedure of deliberate 'data redundancy', care should be taken to ensure that the readings are at least in the right order of magnitude before accepting them.

16.12 BASE STATIONS

Some methods use fields that vary in time. These include the magnetic and gravity fields and natural EM fields. The magnetic field has a daily variation called the 'diurnal' change, the gravity field varies due to tidal effects of the moon, and EM fields vary due to solar phenomenon. To correct for these variations, the field is either monitored separately or readings are repeated at specific time intervals.

16.12.1 CONTINUOUS RECORDING

The best way to detect and monitor temporal variations is to have a second instrument placed at some fixed position with continuous recording capability. This is particularly necessary in the case of magnetic surveying, when rapid changes in the field due to solar activity cause magnetic 'storms'. This type of recording is then often called a 'storm monitor'. A danger with a recording monitor when it is left unattended is that it may malfunction, causing correction information to be lost. To avoid this as much as possible, it should be checked periodically.

16.12.2 PERIODIC REPEAT READING

If a second instrument is not available to be used as a monitor, then field changes can be measured with the one survey instrument by taking repeat readings at a fixed position over time. This can also be done to determine the amount of 'drift' in an instrument and to correct for it.

16.12.3 GRADIOMETERS

The use of gradiometers, where two or more sensors are read at the same time, avoids the need for base monitoring. Since a gradiometer involves the measurement of a field over short spacings simultaneously, any temporal change is the same at these places, and when a gradient is obtained by calculating the difference in two readings, its effect is cancelled out. It is especially advantageous for airborne surveying. Gradiometers also remove any common values which may be unwanted 'regional' effects.

16.13 SIGNAL AND NOISE

Field operations are very much concerned with these two factors, and it is important to understand their significances and differences.

16.13.1 DEFINITIONS AND EXAMPLES

Signal is the reading that is measured by the instruments. It can be in two main parts—the anomaly that is wanted and an unwanted part added to it that is described by the noise. Variable noise (varies with time and/or position, for example, not coherent) are natural atmospheric electromagnetic signals and vibrations due to wind or traffic. Noise can also be a consistent (background response) and not providing information on the objective of the search, for example, power line field or the magnetic response from a nearby pipeline or cable.

However, interfering noise can, for other methods, be the wanted signal or the source field for the measurements. The atmospherics that might disturb an electrical or EM survey can be the source field for another type of EM method (magnetotellurics).

These concepts of signal and noise can be considered by forming the ratio as S/N. If this ratio has a high value, it can indicate that signal is high or noise is low or both. Because the presence of noise can be so common, it is recognized that, in theoretical modelling of synthetic targets, some noise has to be artificially added to make the results more realistic. Often, for want of a better idea, random Gaussian noise of some small percentage of the total signal is added.

By adding a number of readings (stacking), the noise may be cancelled or reduced. Some instruments have ‘filters’ for rejecting predictable noise. A common filter for EM methods is the power line frequency (50 Hz).

16.14 DATA SHEET FORMAT FOR COMMON GEOPHYSICAL METHODS

Besides recording the geophysical data in the instrument as applicable, it is a good practice to keep a record of data in the field notebook also. The standard format for data recoding in the field notebook is provided in the following for common geophysical methods.

16.14.1 WENNER ELECTRICAL RESISTIVITY DATA SHEET

Station: Project: Operator: Remarks:		Direction: Location: Equipment:			
		$\rho_a = 2\pi a \frac{\Delta V}{I}$			
		Date:			
a (m)	V (mV)	I (mA)	K	ρ_a (Ohm-m)	Remarks
50	—	—	—	—	—
—	—	—	—	—	—
100	—	—	—	—	—
—	—	—	—	—	—
200	—	—	—	—	—
—	—	—	—	—	—

16.14.2 SCHLUMBERGER ELECTRICAL RESISTIVITY DATA SHEET

Station:		Direction:					
Project:		Location:					
Operator:		Equipment:					
Remarks:		$\rho_a = \pi \frac{\left(\frac{AB}{2}\right)^2 - \left(\frac{MN}{2}\right)^2}{MN} \frac{\Delta V}{I}$				Date:	
$\frac{AB}{2}$ (m)	$\frac{MN}{2}$ (m)	I (mA)	V (mV)	K	ρ_a (Ohm-m)	Remarks	
1	0.2						
1.5	0.2						
2	0.2						
3	0.2						
5	0.2						
5	1						
7	1						
10	1						
10	2						
12	2						
15	2						
20	2						
25	2						
25	5						
30	5						
40	5						
50	5						
50	10						
—	—	—	—	—	—	—	

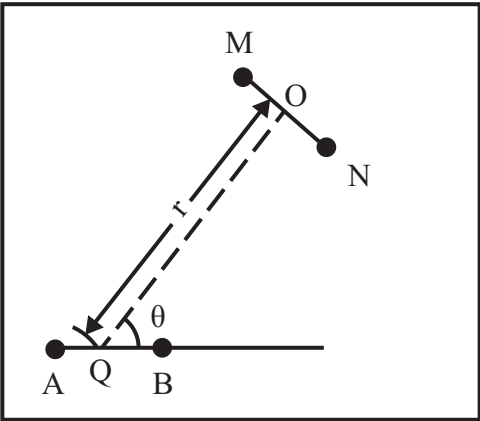
16.14.3 COLLINEAR DIPOLE–DIPOLE ELECTRICAL RESISTIVITY DATA SHEET (TRAVERSING)

Station:		Direction:		Date:				
Project:		Location:						
Operator:		Equipment:						
Remarks:		$\rho_a = \pi a n(n + 1)(n + 2) \frac{V}{I}$						
T_x	R_x	n	a (m)	V (mV)	I (mA)	K	ρ_a	Remarks
1–2	3–4	1	10					
2–3	4–5	1	10					
3–4	5–6	1	10					
—	—	—	—	—	—	—	—	—
1–2	4–5	2	20					

Station:			Direction:				Date:	
Project:			Location:					
Operator:			Equipment:					
Remarks:			$\rho_a = \pi a n(n + 1)(n + 2) \frac{V}{I}$					
T_x	R_x	n	$a \text{ (m)}$	$V \text{ (mV)}$	$I \text{ (mA)}$	K	ρ_a	Remarks
2-3	5-6	2	20					
—	—	—	—	—	—	—	—	—
1-2	5-6	3	30					
2-3	6-7	3	30					
3-4	7-8	3	30					
—	—	—	—	—	—	—	—	—

16.14.4 DIPOLE–DIPOLE ELECTRICAL RESISTIVITY DATA SHEET (SOUNDING, AZIMUTHAL TECHNIQUE)

It can be shown that the electric field, due to a dipole at a given point, is inversely proportional to the cube of the distance r and that, for a given azimuth angle θ , the value of the apparent resistivity ρ_a is a function of the single distance variable r .



Station:			Direction:			Date:	
Project:			Location:				
Operator:			Equipment:				
Remarks:			$\rho_a = K_{\text{azimuthal}} \frac{V}{I} = \frac{2\pi r^3}{L \sin\theta} \frac{V}{I}$				
r (m)	$\frac{AB}{2}$ (m)	$\frac{MN}{2}$ (m)	V (mV)	I (A)	sinθ	K	ρ _a (Ω – m)
600	100	20					
700	100	20					
800	100	20					
1,000	100	20					
1,200	100	20					
1,500	100	20					
1,500	250	50					

(Continued)

Station:			Direction:			Date:	
Project:			Location:				
Operator:			Equipment:				
Remarks:			$\rho_a = K_{\text{azimuthal}} \frac{V}{I} = \frac{2\pi r^3}{L \sin\theta} \frac{V}{I}$				
r (m)	$\frac{AB}{2}$ (m)	$\frac{MN}{2}$ (m)	V (mV)	I (A)	sinθ	K	ρ _a (Ω – m)
2,000	250	50					
2,500	250	50					
3,000	250	50					
3,000	500	100					
4,000	500	100					
5,000	500	100					
6,000	500	100					
6,000	1,000	200					
8,000	1,000	200					
10,000	1,000	200					
—	—	—	—	—	—		

16.14.5 DIPOLE–DIPOLE ELECTRICAL RESISTIVITY DATA SHEET
(SOUNDING, EQUATORIAL TECHNIQUE)

When the azimuth angle θ formed by the line r and the current dipole AB equals $\frac{\pi}{2}$, the azimuthal array and the parallel array reduce to the equatorial array (Figure 16.1). The apparent resistivity remains a function of the single distance variable, $\bar{R} = \sqrt{\left(\frac{AB}{2}\right)^2 + r^2}$.

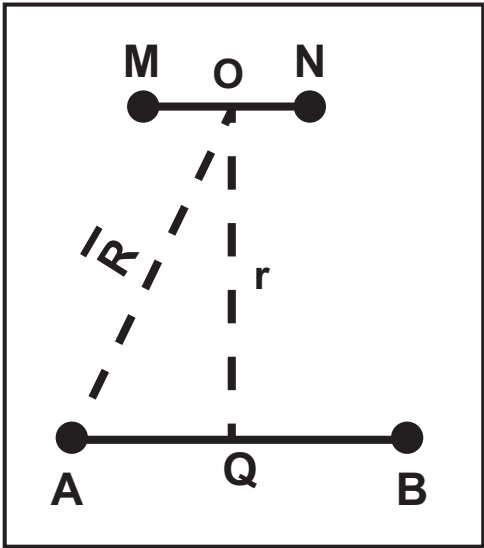


FIGURE 16.1 Dipole–dipole equatorial array.

Station:			Direction:			Date:	
Project:			Location:				
Operator:			Equipment:				
$\rho_a = K_{eq} \frac{V}{I} = \frac{2\pi r^3}{LI} \frac{V}{I}$							
r (m)	$\frac{AB}{2}$ (m)	$\frac{MN}{2}$ (m)	V (mV)	I (A)	K	ρ_a (Ω-m)	Remarks
600	100	20					
700	100	20					
800	100	20					
1,000	100	20					
1,200	100	20					
1,500	100	20					
1,500	250	50					
2,000	250	50					
2,500	250	50					
3,000	250	50					
3,000	500	100					
4,000	500	100					
5,000	500	100					
6,000	500	100					
6,000	1,000	200					
8,000	1,000	200					
10,000	1,000	200					
—	—	—	—	—	—		

16.14.6 SELF-POTENTIAL (SP) DATA SHEET

Area:		Date:				
Voltage Scale: 10 mV full scale		Line:				
Distance from Line Base (m)	Normal Voltage (mV)	Reversed Voltage (mV)	Base Roving Pot Drift Voltage (mV)	Drift Correction	Base Tie in Correction	Absolute Voltage (mV)
0			−2.0	+2	−30	−30
10	−10.1	+10.2		+2		−38
20	−5.3	+5.8		+2		−33
30	−2.1	+1.7		+2		−30
40	+1.5	−1.0		+2		−28
50	+1.8	−3.0		+1		−27
60	+3.5	−5.5		+1		−25
70	+12.4	−10.9		+1		−17
80	+27.8	−26.9		+1		−1
90	−50.1	+52.3		0		−80
100	−80.2	+81.5		0		−110
—	—	—	—	—	—	—

16.14.7 INDUCED POLARIZATION (IP) DATA SHEET

There are several types of IP techniques, but essentially, IP is an extension of traditional resistivity surveying. The ground is characterized not only by its resistivity but also by its chargeability

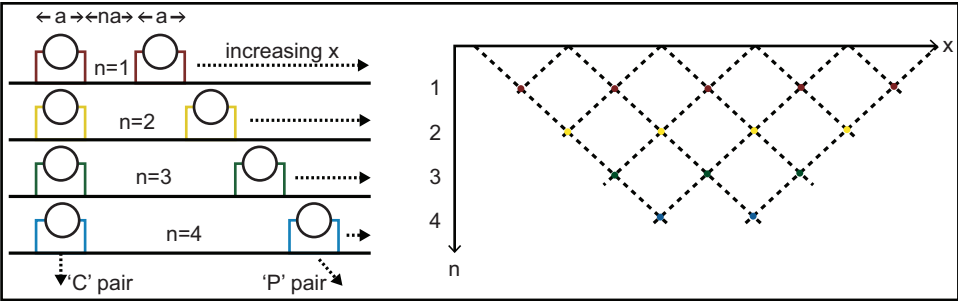


FIGURE 16.2 The dipole–dipole array. *Left:* electrode positions on the ground. *Right:* Plotting positions for each x , a , n combination.

to understand how well it holds electrical charge. The same configurations of electrodes, or arrays, can be used as in resistivity surveying, but the most popular are the dipole–dipole and pole–dipole arrays (Figure 16.2). The depth of investigation for any electrode spacing a is increased by using greater electrode pair separation n values.

Station:			Direction:			Date:		
Project:			Location:					
Operator:			Equipment:					
Remarks:								
			$\rho_a = \pi a n(n + 1)(n + 2) \frac{V}{I}$					
T_x	R_x	n	a (m)	V (mV)	I (mA)	M (mV/V)	K	ρ_a (Ω –m)
1–2	3–4	1	10					
2–3	4–5	1	10					
3–4	5–6	1	10					
—	—	—	—	—	—	—	—	—
1–2	4–5	2	20					
2–3	5–6	2	20					
—	—	—	—	—	—	—	—	—
1–2	5–6	3	30					
2–3	6–7	3	30					
3–4	7–8	3	30					
—	—	—	—	—	—	—	—	—

16.15 DETERMINING THE ANOMALY

The purpose of geophysical surveying can be simply stated as the search for an anomaly or anomalies. *Anomalies* are variations or changes from ‘normal’ values, also referred to as ‘background’, and the difference from background is the quantitative measure of the anomaly. To determine the value or spatial extent of an anomaly requires the background to be defined. Some spatial background variations are known theoretically, such as the normal gravity and magnetic fields (the latter being described by the IGRF). Even after the removal of these standard variations, there may still be a smoothly varying background field existing, called regional, due to deeper geological formations than the zone of interest. When it is removed from the readings to obtain the anomaly, what remains is referred to as the ‘residual’. Gradiometers also remove deep-seated responses and so act as a filter to regional variations.

16.16 LIST OF SELECTED GLOSSARY, ACRONYMS, SYMBOLS, AND TERMS

Absorption	The process of removal of a chemical or other substance from a fluid by attachment of the species to solid surfaces in the transmitting material.
Acoustic	An adjective referring to properties or devices that depend on pressure waves in or near the audible range of about 20–20,000 Hz.
Acoustic velocity	Refers to the speed with which seismic waves propagate through a medium. Acoustic velocity is a function of the engineering properties of the medium and is not a vector quantity when used in this sense.
Acoustic impedance	A seismic wave parameter that is the product of density and seismic wave speed ($Z = \text{density} \times \text{speed} = \rho v$).
Acoustic log	A well log that displays a continuous record of the seismic velocity of the rocks in the borehole wall as a function of depth; also called sonic log or continuous velocity log.
Acoustic wave	A propagating pressure wave, frequently used as an approximation to a full seismic wave.
Air wave	An acoustic wave that propagates through the air from a seismic source.
Alkali vapour magnetometer	A magnetometer that relies on the dependence of energy levels in alkali atoms on the magnetic field.
Amplitude spectrum	The distribution of the amplitudes of the Fourier components of a signal as a function of frequency.
Attenuation coefficient	A numerical measure of the rate of attenuation with distance.
Bandwidth	The range of frequency content of a signal or the range of frequencies passed by a system.
Bistatic mode	In ground-penetrating radar: a technique which uses separate transmitting and receiving antennae.
Borehole log	A graphical record of the materials and/or their properties as a function of depth in a borehole.
Brownfields	Real property whose redevelopment may be complicated by the presence or potential presence of pollution.
Calliper log	A device for logging a borehole which measures the diameter of the hole continuously along its length.
Capacitivity	A measure of permittivity, the property of a material which enables it to store electric charge.
Caesium magnetometer	A devise for measuring the Earth's magnetic field which utilizes the magnetic field dependence of the atomic energy levels in caesium.
Chargeability	A measure of the time domain induced polarization related to secondary to primary voltage ratios.
Cole–Cole	A method of characterizing the time dependence of induced polarization effects using an equivalent circuit.
Common midpoint (CDP)	A technique of seismic reflection data acquisition where traces having different ray paths but the same midpoint between source and receiver are summed after adjustments for geometry.
Complex conductivity	The ratio of the current to the voltage when they are not in phase, used in induced polarization methods.
Complex resistivity	The ratio of the voltage to the current when they are not in phase, used in induced polarization methods.
Compressional waves	A longitudinal mechanical wave that propagates compressions and dilations of the material.
Convolution	A mathematical procedure for determining the output of a linear system for a specified input signal.
Critical angle	When a wave strikes an interface, where the wave propagation velocity increase.
Critical distance	In seismic refraction, the distance from the source at which reflection and refraction times are the same; this corresponds to the onset of the head wave arrival.

Cross-hole tomography	A technique for determining the distribution of physical properties of the zone between two boreholes by making an extensive series of measurement with the source(s) in one hole and the receiver(s) in the other; this applies to many techniques, such as resistivity, seismic, radar, etc.
Crossover distance	In seismic refraction, it is the distance at which the head waves from the first refracting layer arrive at the same time as the direct waves.
Curie temperature	The maximum temperature at which a ferromagnetic material can remain ferromagnetic; variable, but in the neighbourhood of 550°C for a number of typical materials.
Current density	The current flow per unit area through a small surface area, measured in amperes/square metre.
Deconvolution	The mathematical procedure for removing the response of a linear system from the output signal to find the input signal.
Depth of investigation	The maximum depth at which a geophysical technique provides useful information about the subsurface; it depends on the technique, the local conditions, and the type of target.
Depth of penetration	The depth to which currents or fields used for exploration penetrate the subsurface, closely related to depth of investigation.
Dielectric constant	The dielectric constant of a material is a measure of its capacity to store charge when an electric field is applied.
Dilatational wave	A seismic wave or other mechanical wave where the particle motion is along the line of wave propagation; also called <i>P</i> waves and compressional waves.
Dispersion	Spreading of a material or field in space; this term has many uses, including molecular dispersion, spreading of a wave form due to velocity dependence on wavelength, transient electric field spreading in the subsurface, etc.
Diving wave	A seismic wave that is continually refracted as the ray propagates through a strong velocity gradient medium so the wave bends and eventually returns to the surface.
Elastic moduli	Acoustic velocity is a function of the elastic moduli of the medium through which it is propagating. These moduli include bulk modulus, shear modulus, Young's modulus, Lamé's constant, and Poisson's ratio.
Elastic wave	A wave that propagates by elastic deformation of the transmitting material.
Electrical conductivity	A measure of the ability of a material to conduct electric current.
Electromagnetic (EM) velocity	Refers to the speed with which electromagnetic waves propagate through a medium. Electromagnetic velocity is a function of the dielectric property of the medium through which it is propagating and is not a vector quantity when used in this sense.
Equipotential surface	A theoretical surface which is the locus of all points having the same value of potential, most commonly used for electric potential.
Equivalence	In geophysical interpretation, this is the range of values for a combination of parameters such that the resulting response difference is too small to confidently detect in a practical situation.
Geophysics	The study of the Earth by quantitative physical methods.
Inductivity	A measure of magnetic permeability.
Magnetic permeability	The magnetic permeability of a material is the ratio of the induced magnetic field to the inducing field.
Magnetic susceptibility	A measure of the degree to which a material can be magnetized.
Natural electric potentials	Voltages that are caused by natural processes, such as flowing water or oxidization and reduction.
Remanent magnetization	The permanent magnetization within rocks. This is caused by natural causes, such as cooling through the Curie point, chemical processes, alignment of magnetic mineral particles during deposition, pressure and exposure to an external magnetic field.
AB	Distance between transmitter electrodes in resistivity/IP surveys.
a, na	Dipole length, number of dipole lengths.
Chargeability	IP response in time domain IP.
EM	Electromagnetic.
Frequency domain	Measuring with continuous single-frequency signal.

GPS	Global Positioning System.
GIS	Geographic information system.
HI	Height of instrument (in levelling, and in gravity survey).
HLEM	Horizontal loop electromagnetic.
HEM	Helicopter electromagnetic.
Hz	Hertz, measurement of frequency (1 Hz = 1 cycle per second).
IP	Induced polarization.
SP	Self-potential.
K	Potassium.
KHz	Kilohertz = 1,000 Hz.
MeV	Million electron volts (measure of energy for nuclear radiation).
mrad	Milliradian (unit of measurement of IP response in phase-angle IP).
mS	Millisiemen (1,000 mS = 1 S).
mS/m	Millisiemen per metre (1/(1 mS/m) = 1,000 Sm).
mGal	Milligal (1 mGal = 1×10^{-6} of the Earth's gravitational field).
mV/V	Unit of chargeability in time domain IP.
MN	Distance between receiver electrodes in resistivity/IP surveys.
NaI	Sodium iodide, crystal used in γ ray spectrometers.
nT	Nanotesla, used to be called gamma (γ). The Earth's magnetic field ranges between approximately 25,000 nT (at equator) and 60,000 nT (at pole).
G	Gauss (1G = 10^5 nT).
PFE	Percent frequency effect (unit of measurement of IP response in frequency domain IP).
REE	Rare Earth elements.
S	Siemen (unit of electrical conductivity).
Th	Thorium.
Tl	Thallium.
Time domain	Measuring with repeated broadband impulses.
TDEM, TEM	Time domain electromagnetic.
U	Uranium.
VLF	Very low frequency (15–25 KHz), type of EM survey.
α , β , γ	Alpha, beta, gamma (types of nuclear radiation).
μ	Magnetic permeability.
Ω -m	Ohm-metre (unit of resistivity).
ρ	Resistivity (units of Ω -m).
σ	Conductivity (units of mS/m).
σt	Conductivity thickness product (units of mS).
ω	Angular frequency in radians/second.

REFERENCES

- Breiner, S., 1973. *Applications Manual for Portable Magnetometers*. San Jose, CA: Geometrics. 2190 Fortune Drive, San Jose, California 95131, U.S.A.
- Clark, S.P. (ed), 1966. *Handbook of Physical Constants. Geological Society of America Memoir 97*. Revised edn. New York: Geological Society of America.
- Parasnis, D.S., 1986. *Principles of Applied Geophysics*. 4th edn. New York: Chapman and Hall.
- Raiche, A.P., Jupp, D.L.B., Rutter, H. and Vozoff, K., 1985. The joint use of coincident loop transient electromagnetic and Schlumberger sounding to resolve layered structure. *Geophysics*, vol. 50, pp. 1618–1627.
- Telford, W.M., Geldart, L.P., Sheriff, R.E. and Keys, D.A., 1976. *Applied Geophysics*. New York: Cambridge University Press.
- Zhdanov, M.S. and Keller, G.V., 1994. *The Geoelectrical Methods in Geophysical Exploration*. Amsterdam: Elsevier.



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Index

Note: Page numbers in *italics* indicate figures, and page numbers in **bold** indicate tables in the text

#

- 1-D Earth, 328
- 2-D and 3-D electrical resistivity tomography, 194–198
 - electrical resistivity tomography field survey design and data acquisition, 195–197
 - electrical resistivity tomography inversion, 197–198
- 2-D Earth, 328–329
- 2-D electric imaging, 194
- 2-D gravity model, 85, 94–95, 95
 - Bouguer anomaly profile and, 96
- 3-D galvanic distortion and decomposition, 525
- 3-D seismic reflection surveying, 478
- 3-D configuration of kimberlite bodies, 200–201
- 3-D Earth, 10, 329–330
- 3-D electric imaging, 194
- 3-D galvanic distortion, 333
- 3-D geological models, 10
- 3-D magnetic inversion, 16
- 4-D gravity survey, 5, 203, 298

A

- absolute gravity, measurement of
 - free fall method, 58
 - pendulum method, 57–58
 - rise-and-fall method, 59
- absorption, 295, 298, 481, 539
 - scattering and, 294
 - velocity and, 290
- accelerated weight, 365–366
- acceleration due to gravity
 - due to altitude, 55, 56
 - due to depth, 55, 56, 57
- accountable error, 532
- accuracy, 531
- acidic groundwater, 223
- acoustic, 539
- acoustic impedance, 539
- acoustic log, 539
- acoustic velocity, 539
- airborne geophysics, 2, 14, 425, *see also* geophysics
 - access, 426
 - airborne electromagnetic method
 - audio frequency magnetics (AFMAG), 431, 432
 - curves to model AFMAG and ZTEM, 434
 - development and description of ZTEM system, 432–434
 - estimation of depth of exploration, 434
 - OCCAM 2-D MT software, 434–435
 - airborne magnetic method, 435–437
 - airborne radiometric method
 - analysis and interpretation, 441
 - applications, 438–439
 - background corrections, 440–441
 - detector height, 440
 - instrumentation, 439
 - man-made radioactive isotopes, 438
 - overview, 438
 - radiometric assaying, 440, 440
 - site geometry, 439–440
 - statistics, 439
 - airborne radiometric surveys, 14
- Airborne Visible-Infrared Imaging Spectrometer-Next Generation (AVIRIS-NG), 444
- airguns, 366–367
- air surveys, 65
- air wave, 539
- Airy-Heiskanen model, 73, 74
- alkali vapour magnetometer, 539
- alpha decay, 394, 394–395, 397
- alpha radiation, 398
- Alpine rock glaciers, geophysical imaging of, 202–203
- Ampere's law, 264
- amplitudes/attenuation, 294
 - electromagnetic (EM) induction, 521
 - spectrum, 539
- analytic signal, 136–137, 137
- Andaman and Nicobar Islands
 - crustal research, 348
 - GPS survey by GSI, 458
- anisotropic half-space, 327
- anomaly, *see also* gravity anomalies; magnetic anomalies; modelling gravity anomalies; *specific types*
 - over extended conductor, 274
 - transient self-potential, 223–224
- antenna characteristics, 295–296
- anticline, 89, 89–90
- instrumentation, 439
- man-made radioactive isotopes, 438
- overview, 438
- radiometric assaying, 440, 440
- site geometry, 439–440
- statistics, 439
- Earth's gravity field, 429–430
- Earth's surface topography, 430
- EM system, 425, 426
- Geological Survey of India (GSI), 430
- GPS system, 427–428
- interdisciplinary studies, 426–427
- magnetic observations, 130
- precise position capabilities, 427–428
- requirements, 427–428
- sampling, 426
- Widgie Townsite nickel prospect, Western Australia, 435

antiferromagnetism, 120
 apparent chargeability, 246
 apparent resistivity, 185, 190, 350
 and apparent chargeability, 246, 253
 common shapes of, 191, 517
 computation of, 31, 176, 195, 196
 curves, 191
 Lee-partitioning, 175
 for 1-D Earth, 328
 and phase of EM wave, 522–523
 pseudodepth section along Purnea-Dalkhola-Sevoke, 201
 pseudosections, 245
 variables, 172
 archaeological geophysics, 304–309
 archaeological research, Kreuttal, Austria, 146–147
 archaeomagnetism, 150
 Archie's law, 164, **165**
 Assam Valley by GSI, seismological data, 387
 astatized types of gravimeter, 59–61, 60
 atomic charge, 161
 atoms, 391–392
 attenuation coefficient, 294, 521, 539
 attractive force, acting between two bodies, 47
 audio frequency magnetics (AFMAG), 431–434
 audio magnetotellurics (AMT), 273, 321, 338, 339, 343

B

Baima, China, metal pipes and cables in, 250–251
 band-pass filter, 135
 bandwidth, 292, 296–297, 450, 539
 Bangladesh
 iron distribution contour map, 37
 Jessore Sadar, 35–37
 base metal mineralization
 Karoi-Rajpura, Rajasthan, 253
 Tamil Nadu, India, 255
 bathymetric side-scan sonar, 477
 beta-minus decay, 395
 beta-plus decay, 396, 397
 beta radiation, 398
 BIFROST Project (1996), Europe, GPS survey, 460
 biogenic sediments, 473, *see also* sediments
 Birbhum Coalfield Area, West Bengal, mineral exploration, 344
 bistatic antenna, 287
 bistatic mode, 539
 blue economy, deep-sea nodules exploration, 484–485, **485**, 486
 Bodensee Gravimeter Geosystem GmbH (BGGS), 479
 body waves, 359–360, 360
 compressional waves (*P* waves), 359, 360
 shear waves (*S* waves), 359, 360
 borehole(s)
 geophysical methods, **2**
 gravity on land, measurement of, 63–64, 64
 image logs, 413
 borehole logging, 422–423, 539
 calliper log, 419
 case study, 422–423
 coring process, 422
 corrosion well logging, 421
 cost-benefit ratio, 410

electrical logging
 borehole image logs, 413
 resistivity log, 412
 lithology logs
 gamma ray logging, 417–418
 self/spontaneous potential log, 418–419
 spectral logging, 418
 logging while drilling (LWD), 420–421
 memory log, 421
 mud logging, 422
 nuclear magnetic resonance (NMR) logging, 420
 overview, 409
 porosity logs
 density logging, 413–414
 neutron porosity, 414–416, 415
 sonic logging, 416–417
 self/spontaneous potential log, 418–419
 spectral noise logging (SNL), 420
 techniques and applications, **410**
 wireline logging
 history of, 411–412
 memory data, 411
 real-time data, 411

boring, 9–10
 Bouguer anomaly (BA), 72, 82, 490, 502
 map, 101
 Bouguer correction, 69, 69–70
 Bouguer equation, 77
 Bouguer gravity, 79
 anomalies, 108
 anomaly map of India, 104
 brownfields, 539
 bulk modulus, 359
 Bureau De Recherché Geologies' Et Meniere's (BRGM), 445, 465–466
 buried volcanic structures, Gulf of Naples, Italy, 491–492, 492

C

caesium magnetometer, 151, 539
 calliper log, 419, 539
 capacitvity, 539
 carrier phase measurements, 428
 case studies, *see also* surveys and case histories, marine geophysics
 airborne geophysics
 komatiite-hosted nickel sulphides, Kambalda district, Western Australia, 436–437
 Widgie Townsite nickel prospect, Western Australia, 435
 borehole logging, 422–423
 direct current (DC) resistivity method, 198–205
 geological outcrop within Kambalda district, 422–423
 geophysical imaging of Alpine Rock Glaciers, 202–203
 by cross-hole radar survey, Muragl, 304
 by georadar survey, Murtèl, 303–304
 Murtèl, 286–287
 for geophysical interpretation, 11–12, 12
 global positioning system (GPS), 458–462
 Andaman and Nicobar Islands (2005–2008), 458
 BIFROST Project (1996), Europe, 460

- coseismic crustal deformation, Hokkaido-Toho-Oki Earthquake, Japan, 458, **459**
- coseismic displacements of the 1994 Northridge, California, earthquake, 458
- gravitational sliding, Mt Etna volcano, 460–462
- plate motion models, 459–460
- rapid deformation of the South Flank of Kilauea volcano, Hawaii, USA, 460
- Sanriku-Haruka-Oki earthquake, Japan, 459
- gold exploration in Meyduk-Latala Area (Kerman Province), 256
- GPR applications
 - Ahichchhatra, Uttar Pradesh, India, 306
 - in Australia and Greece, 305
 - and DGPS method to characterize the Litang Fault, China, 307
 - for groundwater in Granitic Terrain, Hyderabad, India, 308–309
 - at the Harappan Site of Dholavira, in Kachchh, Gujarat, 306–307
 - for historic cemetery of Alabama, 306
 - Ladang Sungai Batu, Merbok, Kedah, 305–306
 - in locating fresh water, at Vizag Steel Plant (VSP), Andhra Pradesh, India, 307–308
- gravity anomaly of chromite deposit body, Cuba, 98, **99**
- gravity method in archaeological prospection in Jordan, 102, *103*
- gravity methods in internal characterization of landfills, 96–98, *97*
- gravity survey in Devon Area, England, 99–100, *100*
- gravity surveying, 93–103
- hydrogeological, 34–37
- induced polarization (IP), 248–258
- inverse modelling, Cerro Prieto, Baja California, Mexico, 222
- IP anomaly, Durkin Shoot Nickel Prospect, Kambalda Dome, Western Australia, 256
- laboratory experiments, 251, *251–252*
- Lesser Himalayas, Purnea–Sevoke Transect, 95, 96
- magnetic exploration, 145–151
- magnetic surveying
 - archaeological research for landscape studies, Kreuttal, Austria, 146–147
 - archaeological site characterization of the Jordan Area, Western Asia, 150
 - geospatial archaeology, Lothal Revisitation Project, Gujarat, India, 150–151
 - geotechnical geophysics non-uniqueness, 147
 - magnetic anomaly above Olivine Diabase Dyke, Pishabo Lake, Canada, 149, *149*, **150**
 - nickel sulphide exploration in the Kambalda District, Western Australia, 147–148
 - Sohagpur, Central India, 146, *146*
- magnetotelluric (MT), 341–352
- mineral exploration in small-scale mining in Fazenda Santana, Brazil, 256–258, **257**, 258
- Narmada-Son lineament (NSL) in Madhya Pradesh (MP), 94–95, *95*
- possible triassic folding below Abu Rassain, Western Desert, Iraq, 100–102
- remote sensing techniques, 451–454
 - aerial archaeology, Hornsburg in Lower Austria, 452–453
 - Cu mineralization in Alwar basin, Rajasthan, India, 451–452
 - locating archaeological sites in India, 453, *454*
 - mineral mapping techniques, South-Eastern Rajasthan, 451
 - PTHEM, Aladahalli Area, Hassan District, Karnataka, 452, *453*
 - zagros lithosphere from postseismic deformation of the 2017 m7.3 Kermanshah, Iraq, earthquake, 451
- resistivity method for archaeological site characterization of Jordan, 204
- resistivity method in internal characterization of landfills, Ilhavo, Portugal, 203–204
- Silurian Reef exploration, 490–491
- VES in locating fresh water in coastal terrain at VizagSteel Plant (VSP), Visakhapatnam, Andhra Pradesh, India, 204–205
- Western Freeway Sea Link (WFSL-Phase IIA), 38–39
- Centre for Mathematical Modelling and Computer Simulation (CMMACS), 457
- centrifugal acceleration, 52, *53*
- Chabsar Area, Ahmedabad, Gujarat, geothermal exploration, 344
- characteristic curves, 189
- chargeability, 236, 539, 540
 - apparent, *246*, 253
 - of various materials, 239, **240**
- chemical remanent magnetization (CRM), 121
- Chikmagalur-Kavali, Dharwar Craton, South India, crustal research, 346–347
- circular anomalies, 105
- civil engineering problems, 28–29
- Clairaut's theorem, 499, 500
- coastal regions, 33–34
- cole-cole, 539
 - model for complex resistivity, 238, 238
 - relaxation spectra, 239
- collinear dipole-dipole electrical resistivity data sheet, 534–535
- common midpoint (CDP), 539
- common midpoint (CMP) method, 376, 377, *377*
- complex conductivity, 539
- complex impedance, 237
- complex permittivity of water, 289
- complex resistivity, 539
- compressional waves (*P* waves), 359, *360*, 539
- Compton scattering, 401
- conductance, 159, 188–189
- conduction, process within the Earth, *162*
- conductivity, 160
- contact resistance, 248
- contaminant transport, 32–33
- continental margin, North-West Ireland, 489–490
- convolution, 539
- Coordinated Universal Time (UTC), 455
- coring process, 422
- corrosion well logging, 421
- coseismic crustal deformation, Hokkaido-Toho-Oki Earthquake, Japan, 458, **459**
- coseismic displacements, Northridge, California, earthquake, 458
- cosmogenic sediments, 473
- cost-benefit ratio, 410

coulomb (C), 157
 coupling of electrical circuits, 248
 critical angle, 539
 critical distance, 539
 cross-borehole imaging, 247
 cross-hole (borehole) ERT survey, 196
 cross-hole seismic tomography, **26**
 cross-hole tomography, 540
 crossover distance, 540
 crustal architecture, 103–104
 crustal field, 124
 crustal research, 345–349
 CSIRO Transient Electromagnetic (SIROTEM), 283
 Cuba, gravity anomaly of chromite deposit body, 98, **98**, **99**
 Cuddapah Basin, Southern India, mineral exploration, 344
 cultural coupling effect, 248
 Curie isotherm depth (CID), 487
 Curie temperature, 540
 current density, 540
 geometry for, *169*
 telluric, 226, 227
 current distribution, 167–168, *168–169*
 currents, *see* natural currents

D

data acquisition, 195–197
 and field sites, Risby landfill, 33
 remote sensing techniques, 446
 special techniques, 298
 standard procedure, 297–298
 data enhancement, *79–80*, 79–81
 data format, for reduction of gravity data, 72, **72**
 data interpretation
 electromagnetic (EM), 298
 induced polarization (IP), 241–246, **244**, 245, 246
 magnetic exploration
 qualitative interpretation, 138–139
 quantitative interpretation, 139
 data processing
 magnetic exploration, 134–138
 remote sensing techniques, 449–450
 data space method, 338
 decay of potential, 234
 Deccan Volcanic Province (DVP), Maharashtra, crustal research, 348
 declination, magnetic field, 122
 deconvolution, 81, 540
 deep crustal studies, 201–202
 Deep Ocean Mission, 3
 deep probing case study, 300–301
 deep-sea nodules exploration, blue economy, 484–485, **485**, 486
 Defense Reutilization and Marketing Office (DRMO), 35
 delineation, 103–104, *104*
 of coal-bearing Gondwana rocks, 200
 of Gondwana Sediments below the Deccan trap, 199–200
 of lead (Pb) contamination from, 301–302
 of shallow crustal structure, 199
 density imaging, 77
 density logging, 413–414
 depth of investigation, 540
 depth of penetration, 540
 derivatives, 136
 detector height, 440

detrital remanent magnetization (DRM), 121
 Devon Area, England, gravity survey, 99–100, *100*
 diamagnetism, 119
 Diapir, 85–87, *86*
 dielectric constant, 161–162, **162**, 540
 diffraction tomography, 384–385
 dilatational wave, 540
 dimensionality vs. impedance tensor, **330**
 dipolar magnetic field, 505
 dipole configuration, 216, 217
 dipole-dipole
 arrays, 177–180, *179*, 186, 243, 245
 electrical resistivity data sheet, 535–537, *536*
 sections, *17*, 17–18
 direct current (DC) resistivity method, 157–205
 2-D electrical resistivity tomography, 194–198
 electrical resistivity tomography field survey design
 and data acquisition, 195–197
 electrical resistivity tomography inversion,
 197–198
 2-D electric imaging, 194
 3-D electric imaging, 194
 3-D electrical resistivity tomography, 194–198
 applications and case studies, 198–205
 Archie's law, 164, **165**
 current distribution, 167–168, *168–169*
 electrical principles, 157–164
 atomic charge, 161
 electrical conduction, types of, 161–162, **162**, *162*
 electric field and potential, 158–159, *159*
 equipotentials, 161
 Ohm's law, 159–160, *160*
 potentials in homogeneous media, 162–163
 single-current electrode at depth, 163–164, *164*
 electrode configurations, 174–180
 dipole-dipole arrays, 177–180, *179*, 186
 gradient array, 177, *177*
 Lee-partitioning electrode array, 175, *175*
 pole-dipole array, 176, *176*
 pole-pole array, 176–177, *177*
 Schlumberger array, *175*, 175–176, 186
 Wenner array, *174*, 174–175, 186
 geoelectric parameters, 187–189, *189*
 homogeneous ground, 180–184
 distortion of current flow at plane interface, 180, *180*
 distortion of potential at plane interface, *181–182*,
 181–184
 logarithmic coordinates, 187
 porosity of rocks, **165**
 principles of, 169–174, *170–173*
 probing depth, 187
 resistivity surveying, 185–186
 rock resistivities, 165, **166**, *167*
 vertical electrical sounding, 189–194
 ambiguities of sounding curve interpretation,
 192–194
 over horizontally stratified media, 189–192, *191*
 over laterally inhomogeneous media, 192
 direct interpretation gravity anomalies, 82, **83**
 dispersion, 540
 diurnal variations, 126, 127, 131–132, 507
 diving wave, 540
 The Doppler observable, 428
 Dose channel, 14
 downhole electromagnetic (DHEM), 423

downward continuation, 135
 drift correction, 68
 drift variations, 131–132
 drop-weight, 365–366
 dynamical ellipticity, 52

E

earthquake magnitude, 512–513
 Earthquake precursor prediction research, 349–352
 earthquake precursors, **405**, 405–407, 406
 eddy currents, 264
 elastic moduli, 540
 elastic wave, 540
 elastic wave velocity, 361
 electrical anisotropy, 193
 electrical conduction, types of, 161–162, **162**, 162
 electrical conductivity, 540
 lateral changes of, 272
 electrical logging
 borehole image logs, 413
 resistivity log, 412
 electrical methods, 6–8, 528
 direct current resistivity method, 7
 electromagnetic (EM) methods, 7
 induced polarization method, 8
 mise-a-la-masse method, 7
 self-potential method, 8
 electrical principles, 157–164
 atomic charge, 161
 electrical conduction, types of, 161–162, **162**, 162
 electric field and potential, 158–159, 159
 equipotentials, 161
 Ohm's law, 159–160, 160
 potentials in homogeneous media, 162–163
 single-current electrode at depth, 163–164, 164
 electrical resistivity, **26**, 30–36, 160, 167, 194–198
 electrical resistivity tomography (ERT), 31, 287–289
 inversion, 197–198
 in locating fresh water, 204–205
 survey
 cross-hole (borehole), 196
 mobile, 197
 underwater, 197
 electrical sounding, vertical, *see* vertical electrical sounding
 electric field and potential, 158–159, 159, 513
 electric field lines, 170, 172
 electric properties
 effect of inhomogeneous ground, 515–516
 electrode configurations, 514–515
 field and potential, 513
 force of attraction, 513
 Ohm's law, 513–514
 resistivity method, 514
 resistivity surveying, 516–518
 electrochemical potential, 218
 electrode configurations, 174–180, 514–515
 dipole-dipole arrays, 177–180, 179, 186
 gradient array, 177, 177
 Lee-partitioning electrode array, 175, 175
 pole-dipole array, 176, 176
 pole-pole array, 176–177, 177
 Schlumberger array, 175, 175–176, 186
 Wenner array, 174, 174–175, 186

electrode polarization, 235, 236, 248
 electrolytic/ionic conduction, 161
 electromagnetic (EM), 262–311
 applications for, **263**
 archaeological geophysics, 304–309
 clay horizon through, 272
 electromagnetic induction, 264–266, 265
 ground conductivity meters, 269–273
 horizontal loop EM (HLEM), 267–269
 geophysical imaging of Alpine Rock Glaciers
 by cross-hole radar survey, Muragl, 304
 by georadar survey, Murtèl, 303–304
 geotechnical studies, 309
 ground-penetrating or probing radar (GPR), 287–300
 amplitudes/attenuation, 294
 applications of, 298–300, 299, 305–309
 complex permittivity of water, 289
 data interpretation, 298
 deep probing case study, 300–301
 delineation of lead (Pb) contamination from, 301–302
 electric permittivity and conductivity, 287–289
 electromagnetic wave propagation, 289–290
 exploration depth, 294–295
 handheld step-frequency using fieldfox network analyzer, 302–303
 horizontal and vertical resolution, 292
 ray geometry and travel times, 293–294
 reflection and refraction of plane waves, 291, **292**
 scattering and diffraction, 291–292
 technical aspects of, 295–298
 induction, **26**, 119
 amplitudes/attenuation, 521
 exploration depth, 521
 propagation speed, 520
 reflection and refraction, 520–521
 signal travel times, 521
 principles of, 263–264
 transient electromagnetic (TEM/TDEM) techniques, 279–287
 geophysical imaging of Alpine Rock Glaciers, Murtèl, 286–287
 hydrocarbon exploration, 284–285
 mineral exploration, 282–284
 saltwater intrusion with, 285–286
 using plane wave source of excitation, 273–278
 very-low-frequency (VLF), 273–274, 273–276
 VLF resistivity and radio magnetotelluric (RMT, CSRMT) method, 276–277, 276–278, **278**
 velocity, 540
 electromagnetic coupling, 248
 electromagnetic fields, behaviour of, 322–324
 electromagnetic induction, 264–266, 265
 ground conductivity meters, 269–273
 in homogeneous half-space, 324–327
 horizontal loop EM (HLEM), 267–269
 electromagnetic radiation (EMR), 444
 electromagnetic wave propagation, 289–290
 electron capture, 396–397, 397
 electronics
 conduction, 161
 pulse generators, 296
 receiver systems, 296–297
 elevation correction, 68–71
 ellipsoid, 55
 El Niño-Southern Oscillation (ENSO), 445

elongated anomalies, 105
 EM, *see* electromagnetic
 emissions, 393–394
 engineering applications, **29**
 environmental and engineering geophysics, 25–40
 bearing capacity of gravel boulder deposit from footing
 load test, 37–38
 civil engineering problems, 28–29
 engineering applications, **29**
 environmental applications, **29**
 environmental problems, 28
 geophysical fieldwork, 30
 geophysical methods (*see* geophysical methods)
 geophysical surveying methods for geotechnical
 investigations, **26–27**
 marine geotechnical and geophysical surveys, 38–39
 shallow geophysical investigation, 28
 environmental applications, **29**
 environmental problems, 28
 Environmental Protection Agency (EPA), 36, 198
 Eötvös correction, 71, 482, 502
 Etna volcano, 460–462
 equipotentials, 161, 170, 172
 equipotential surface, 48, 540
 erratic field variations, 131
 excavation, 9–10
 exclusive economic zone (EEZ), 471
 exploration depth, 294–295, 434
 electromagnetic (EM) induction, 521
 exploratory/research vessels, 483
 explosives, 366
 expression potential field, 153n1
 external magnetic field, 124

F

Faraday's law, 263
 ferrimagnetism, 120
 ferromagnetism, 120
 field layout, MT, 339–340
 field lines
 electric, 170, 172
 for magnetic dipoles, 118
 field measurements, 507
 field nomenclature, 506–507
 field procedures, 130–133
 field processing, 241–246, **244**, 245, 246
 field sites and data collection, Risby landfill, 33
 field survey, 241–246, **244**, 245, 246
 seismic reflection method, 375–378
 field variations, erratic, 131
 flow chart
 for interpretational techniques, 83
 for inverse problem, interpretational techniques, 83
 fluxgate magnetometer, 128–129
 focused offshore mineral exploration (FOME), 471
 footing load test, bearing capacity of gravel boulder
 deposit from, 37–38
 forces, 153n6
 associated with magnetic monopoles, 117
 fractured rock aquifer, Naval Air Warfare Centre, New
 Jersey, 34–35
 free air anomaly (FAA), 71, 490, 502
 free air correction, 69, 69

free fall method, 58
 frequency effect (FE), 236–237
 Fresnel zone, 378, 378
 Full waveform inversion (FWI), 361

G

gamma decay, 395, 395–396, 397
 gamma radiation, 398
 gamma ray
 geophysical method, 8–9
 logging, 417–418
 spectrometry, 403
 and X-ray detections, 400
 GANOVEX expeditions, 490
Garuda Vasudha, 445
 Geiger-Muller (GM) counter
 gamma and X-ray detections, 400
 neutron detection, 400
 operating principle, 399–400
 types and applications, 400
 general geophysics, 25
 geochemical datasets, 19–20, **20**
 geochemical method, 9
 geodesy, 429
 geoelectric layer, 188
 geoelectric parameters, 187–189, 189
 geographic information system (GIS), 465
 during constructing process, 463
 geophysical exploration survey, 464–466, **465**
 mineral exploration, 465
 oil and gas exploration, Ethiopia, 464
 seismic/earthquake studies, Aizawl Town, Mizoram, 466
 seismic refraction software, 465
 Tons Valley base metal prospect, Uttar Pradesh (UP),
 465–466
 using ArcGIS, 463
 geoid of reference, 54, 55
 geological datasets, 19–20, **20**
 geological framework, 11
 geological outcrop, Kambalda district, 422–423
 geological structures, 109
 geology and magnetics, 124–126, 125
 geomagnetic data, 151–153
 geomagnetic field of the Earth, 124
 geomagnetic observatories, India, 151–152, **152**
 geometrical parameters, 28
 geophysical anomalies, 2
 geophysical datasets, 19–20, **20**
 geotechnical interpretation of, 18
 geophysical exploration
 of lithium reserves, 18–19
 radioactivity
 buried channel and flow of underground fluids,
 403–405, 404
 minerals and metals, 402–403
 surface gamma activity, 403
 survey, 464–466, **465**
 geophysical fieldwork, 30
 geophysical imaging
 of Alpine Rock Glaciers
 by cross-hole radar survey, Muragl, 304
 by georadar survey, Murtèl, 303–304
 Murtèl, 202–203, 286–287

- geophysical interpretation, case studies for, 11–12, 12
- geophysical methods
- anomalies, 538
 - base metal mineralization
 - Karoi-Rajpura, Rajasthan, 253
 - Tamil Nadu, India, 255
 - base stations, 532
 - characteristics, **496–497**
 - choice of method, 529
 - collinear dipole-dipole electrical resistivity data sheet, 534–535
 - copper and gold mineralization, Rajasthan, 254, 254
 - data sheet format, 533–538
 - definitions, 539–540
 - dipole-dipole electrical resistivity data sheet, 535–537, 536
 - electrical method comparison, 528, **528**
 - electric and electromagnetic properties, 513–521
 - field readings, 531–532
 - force fields, **497**
 - fundamental properties, 495, 496
 - geometrical parameters, 28
 - gravity, 498–504
 - hydrogeological case studies, 34–37
 - hydrologic applications of, 30–31, 30–34
 - induced polarization (IP) data sheet, 537–538, 538
 - instrumentation, 530
 - magnetic properties, 504–507
 - magnetotelluric (natural source low-frequency EM), 521–526
 - marine geophysics, 526
 - mathematical identities, 498
 - multiple methods and survey design, 528–529
 - operational precautions, 530
 - overview, 495
 - of Pb–Zn deposit of Lontzen-Poppelsberg, Belgium, 255–256
 - physical properties of, **527**, 527–528
 - Schlumberger electrical resistivity data sheet, 534
 - seismic waves, 507–513
 - selection of survey parameters, 530–531
 - self-potential (SP) data sheet, 537
 - signal and noise, 533
 - spatial distribution of petrophysical/geophysical parameters, 28
 - surveying for geotechnical investigations, **26–27**
 - symbols and terms, 540–541
 - trial survey, 531
 - Wenner electrical resistivity data sheet, 533
- geophysical modelling, 10–18
- geophysical well logging. *See* borehole logging
- geophysics, 462, 495, 540, *see also* marine geophysics; *specific types*
- defined, 1
 - engineering (*see* environmental and engineering geophysics)
 - exploration surveys and applications, **2**
 - geochemical datasets, 19–20, **20**
 - geological and geophysical datasets, 19–20, **20**
 - geophysical exploration of lithium reserves, 18–19
 - geotechnical interpretation of geophysical data, 18
 - integrated geological and geophysical modelling, 10–18
 - case studies available for geophysical interpretation, 11–12, 12
 - regional basement modelling, 12–18, 15
 - methodology and instrumentation, 4–10
 - boring, 9–10
 - electrical methods, 6–8
 - excavation, 9–10
 - gamma ray geophysical method, 8–9
 - geochemical method, 9
 - geothermal method, 9
 - gravity method, 4–5
 - magnetic method, 5–6
 - radioactive method, 8
 - remote sensing method, 9
 - sampling, 9–10
 - seismic method, 6
 - objectives and accomplishments, 3–4
- Geosoft Program Package, 16
- geospatial archaeology, Lothal Revisitation Project, Gujarat, India, 150–151
- geotechnical geophysics non-uniqueness, 147
- geotechnical interpretation, of geophysical data, 18
- geotechnical studies, 309
- geothermal exploration, Chabsar Area, Ahmedabad, Gujarat, 344
- geothermal method, 9
- German Antarctic North Victoria Land Expedition (GANOVEX) expeditions, 490
- GIS, *see* geographic information system
- global navigation satellite system (GNSS), 457
- global positioning system (GPS)
- accuracies and comparison, 455–456
 - applications of, 456
 - case study, 458–462
 - in India, 456–458
 - overview, 455
 - very-small-aperture terminal (VSAT), 457
- GPR, *see* ground-penetrating or probing radar
- gradient, 105
- gradient array, 177, 177
- gradient configuration, 216, 217
- gradient IP map, 16
- gradient resistivity map, 16
- gradients of gravity, 51
- radiometers, 481, 532
- gravel boulder deposit, bearing capacity of, 37–38
- gravimeter, 25, 57
 - astatized types of, 59–61, 60
 - LaCoste-Romberg, 59
- gravitational acceleration of point masses, 48–50
- gravitational potential of point masses, 48–50
- gravitational sliding, Mt Etna volcano, 460–462
- gravitation attraction, 48
 - of small mass m on Earth, 47
- gravity, **27**
- accuracy, 501
 - anomalies, 503–504
 - base tie, **67**
 - bouguer anomaly (BA), 502
 - comparison of magnetic anomalies with, 139–141
 - data record sheet, **67**
 - Eötvös correction, 502
 - free air anomaly (FAA), 502
 - gradient of gravity, 499
 - gravitational potential, 499
 - isostasy, 502

- isostatic anomaly (IA), 502
 - latitude correction, 502
 - and magnetics, differences between, 116
 - and magnetics, similarities between, 115–116
 - mass of the Earth, 499
 - measurement of absolute gravity, 501
 - measurement of component of, 61–62, 61–63
 - measurement of relative gravity, 501
 - measures only vertical component, 61
 - Newton's law, 499
 - potential field equations, 499
 - resolution of gravimeter, 501
 - rock density, 503
 - second derivative, 499
 - sensitivity, 501
 - shape of the Earth, 499–500
 - solar system data, 504
 - spheroid and geoid, 500
 - surveying, 502
 - terrain correction, 502
 - variation in acceleration, 500
 - gravity anomalies, 71–72
 - across a vertical fault, 93
 - across horizontal thin sheet, 91
 - Bouguer anomaly, 72
 - for buried spheres, 86
 - of chromite deposit body, Cuba, 98, 98, **99**
 - data enhancement, 79–80, 79–81
 - data format for reduction of gravity data, 72, **72**
 - free air anomaly, 71
 - of infinitely long linear mass, 88
 - interpretation for potential fields, 77–93
 - inverse problem, interpretational techniques, 82–85
 - irregular shape, 62, 62–63
 - isostatic anomaly, 72
 - modelling, 85–93
 - point mass, 62, 62
 - of point mass or sphere, 85–87, 86
 - regional-residual gravity anomaly, 77–78, 78
 - gravity data
 - data format for reduction of, 72, **72**
 - of India for delineation of structural fabric and crustal architecture, 103–104, 104
 - reduction for gravity surveying, 67–71
 - gravity effect of a sphere, 83
 - gravity field, 153n3
 - gravity/magnetic survey, 478–481
 - radiometer, 481
 - measured values, 479
 - optically pumped magnetometer, 481
 - overhauser magnetometers, 480
 - proton precession magnetometers, 480
 - gravity maps, of mineral deposits, 105–106, 105–107, **107**, 108
 - gravity methods, 4–5
 - in archaeological prospection in Jordan, 102, 103
 - in internal characterization of landfills, 96–98, 97
 - gravity model, of Ilhavo landfill, 97
 - gravity on land, measurement of
 - in boreholes, 63–64, 64
 - on the Earth's surface, 57–63
 - gravity on moving platforms, measurement of
 - air surveys, 65
 - sea surveys, 64–65
 - space measurements, 65
 - gravity signatures, of mineral deposits, 109, **109–110**
 - gravity surveying, 44, 65–71
 - applications of, **44**, 93–103
 - case study, 93–103
 - in Devon Area, England, 99–100, 100
 - reduction of gravity data, 67–71
 - GRIDEPTH program, 16
 - ground-based magnetic observations, 130–131
 - ground conductivity meters, 269–273
 - ground geophysical survey equipment, 30
 - ground-penetrating or probing radar (GPR), **26**, 30–31, 287–300, 519–521
 - amplitudes/attenuation, 294
 - applications of, 298–300, 299, 305–309
 - Ahichchhatra, Uttar Pradesh, India, 306
 - in Australia and Greece, 305
 - and DGPS method to characterize the Litang Fault, China, 307
 - for groundwater in Granitic Terrain, Hyderabad, India, 308–309
 - at the Harappan Site of Dholavira, in Kachchh, Gujarat, 306–307
 - for historic cemetery of Alabama, 306
 - Ladang Sungai Batu, Merbok, Kedah, 305–306
 - in locating fresh water, at Vizag Steel Plant (VSP), Andhra Pradesh, India, 307–308
 - complex permittivity of water, 289
 - data interpretation, 298
 - deep probing case study, 300–301
 - delineation of lead (Pb) contamination from, 301–302
 - electric permittivity and conductivity, 287–289
 - electromagnetic wave propagation, 289–290
 - exploration depth, 294–295
 - handheld step-frequency using fieldfox network analyzer, 302–303
 - horizontal and vertical resolution, 292
 - ray geometry and travel times, 293–294
 - reflection and refraction of plane waves, 291, **292**
 - scattering and diffraction, 291–292
 - technical aspects of, 295–298
 - utility of, 299
 - groundwater
 - contamination
 - with cyanide complexes, Czechoslovakia, 249, 249–250
 - exploration, Sauk-Soo River, in Crimea and in Armenia, 250
 - resistivity survey, 198–199
 - Guatemalan volcanoes, 221
 - Gulf of Naples, Italy, buried volcanic structures, 491–492, 492
- ## H
- half-life, 392–393, **393**
 - Hamirpur-Rupadia, Indo-Gangetic Plain, crustal research, 347
 - hammer, 365
 - handheld step-frequency using fieldfox network analyzer, 302–303
 - hazard management, baseline geophysical data for, 488–489
 - heat conduction, 264
 - heliborne geophysical survey system (HGSS), 444, 445
 - Helmholtz wave equation, 522

- high-pass filter, 135, *see also* low-pass filter
 - Hokkaido-Toho-Oki Earthquake, Japan, coseismic crustal deformation, 458, **459**
 - homogeneous ground, 180–184
 - distortion of current flow at plane interface, 180, 180
 - distortion of potential at plane interface, 181–182, 181–184
 - homogeneous half-space, electromagnetic induction in, 324–327
 - horizontal and vertical resolution, 292
 - horizontal cylinder
 - modelling gravity anomalies, 89–90, 89–90
 - signal enhancement in case of, 80
 - horizontal derivative, 136
 - horizontal gradient, 79
 - of gravity from top and bottom of tower, 51
 - horizontal interface, 379, 379–381
 - horizontal line element, modelling gravity anomalies, 87–89
 - horizontal loop EM (HLEM), 267–269
 - horizontal resolution, 378
 - horizontal slab model for vertical fault, 92–93
 - horizontal thin sheet
 - gravity anomaly across, 91
 - modelling gravity anomalies, 90–92
 - hotspots, 385–386
 - hydrocarbon exploration, 284–285, 343
 - hydrogen index (HI), 414
 - hydrogenous sediments, 473
 - hydrogeological case studies, 34–37
 - hydrological characterization and controls
 - Hanford site located in Richland, Washington, 34
 - hydrologic applications of geophysical methods
 - field sites and data collection, Risby landfill, 33
 - landfills and contaminant transport, 32–33
 - regional geology and coastal regions, 33–34
 - vadose zone, 30–31, 30–32
 - hyperspectral remote sensing (Hyperion), 444
 - hyperspectral vs. multispectral remote sensing, 450
 - hysteresis loop, 119
- I**
- Ilhavo, Portugal, resistivity method in landfills, 203–204
 - impedance tensors, 327–331
 - decomposition of, 333
 - vs. dimensionality and isotropic conductivity distribution, **330**
 - rotation-invariant magnetotelluric, 333–335
 - rotation of, 330–331
 - inclination, magnetic field, 122
 - inclined interface, 381, 381–383
 - India
 - base metal mineralization, Tamil Nadu, 255
 - Bouguer gravity, anomaly map of, 104
 - Cu mineralization in Alwar basin, Rajasthan, 254, 254, 451–452
 - for delineation of structural fabric and crustal architecture, 103–104, 104
 - geomagnetic observatories, 151–152, **152**
 - gravity data for delineation of structural fabric and crustal architecture, 103–104, 104
 - locating archaeological sites in, 453, 454
 - Lothal Revisitation Project, Gujarat, geospatial archaeology, 150–151
 - magnetotelluric (MT) survey groups in, **342**
 - Mundgod-Siruguppa, Dharwar Craton, South India, crustal research, 346
 - Narmada-Son lineament (NSL), 94–95, 95
 - indirect interpretation of gravity anomalies, 84, 84
 - induced magnetization, mechanisms for, 119–120
 - induced polarization (IP), **26**, 33, 233–258, *see also* magnetic induced polarization
 - anomaly over Durkin Shoot Nickel Prospect, Kambalda Dome, Western Australia, 256
 - applications and case studies, 248–258
 - associated with constriction between mineral grains, 235, 235
 - chargeability of various materials, 239, **240**
 - cole-cole model for complex resistivity, 238
 - cole-cole relaxation spectra, 239
 - cross-borehole imaging, 247
 - data sheet, 537–538, 538
 - decay of potential, 234
 - effect over freshwater and saltwater pockets, 252
 - electrode polarization IP effect, 236
 - field survey, processing, and interpretation of data, 241–246, **244**, 245, 246
 - gold exploration in Meyduk-Latala Area (Kerman Province), 256
 - instruments, 240–241
 - interpretation in environmental investigations, Crescent Beach State Park, Maine, 250
 - limitations of, 247
 - magnetic induced polarization (MIP), 239
 - measurement errors, 247–248
 - phase domain, 237–238
 - polarization of the Earth, 233–237
 - principles of, 233–237
 - induction arrows, 332
 - inductivity, 540
 - inhomogeneous ground, 515–516
 - in-phase component, 266
 - instrumentation/methodology, 4–10
 - airborne radiometric method, 439
 - boring, 9–10
 - electrical methods, 6–8
 - excavation, 9–10
 - gamma ray geophysical method, 8–9
 - geochemical method, 9
 - geothermal method, 9
 - gravity method, 4–5
 - induced polarization (IP), 240–241
 - magnetic method, 5–6
 - magnetic properties, 507
 - magnetotelluric (MT), 339
 - radioactive method, 8
 - remote sensing method, 9
 - sampling, 9–10
 - seismic method, 6
 - instrument drift, 131
 - integrated geological modelling, 10–18
 - case studies available for geophysical interpretation, 11–12, 12
 - regional basement modelling, 12–18, 15
 - integrated interpretation map, geological and geophysical results, 17
 - interference and magnetic cleanliness, 131
 - interferometric sonar, 476–477
 - Intermagnet network, 152

interpretational techniques, flow chart for, 83
 inverse modelling, Cerro Prieto, Baja California, Mexico, 222
 inverse problem, 82–85
 interpretational techniques
 direct interpretation, 82, **83**
 flow chart for, 83
 gravity effect of a sphere, 83
 indirect interpretation of gravity anomalies, 84, 84
 non-unicity of the solution, 85
 inversion routines, magnetotelluric (MT), 336–338
 IP, *see* induced polarization
 irregular shape
 gravity anomalies, 62, 62–63
 schematic diagram of, 62
 schematic diagram of whole body, 62
 isomeric transition, 396
 isostasy
 Airy-Heiskanen model, 73, 74
 isostatic compensation and vertical crustal movements, 74–75
 Pratt-Hayford isostatic model, 73–74, 74
 isostatic anomaly (IA), 72, 502
 isostatic compensation, 74, 74–75
 isostatic correction, 71
 isothermal remanent magnetism (IRM), 121
 isotropic conductivity distribution vs. impedance tensor, **330**

J

Jammu-Udhampur-Katra-Quazigund-Baramulla railway line, 37–38
 Japan
 coseismic crustal deformation, Hokkaido-Toho-Oki Earthquake, 458, **459**
 Sanriku-Haruka-Oki earthquake, 459
 in volcanoes, 220–221
 Jessore Sadar, Bangladesh, 35–37
 Jordan
 gravity method in archaeological prospection, 102, 103
 resistivity method for archaeological site characterization of, 204

K

Kambalda district, Western Australia, komatiite-hosted nickel sulphides, 436–437
 Kepler's Laws of Planetary Motion, 44–46, 45–46

L

LaCoste-Romberg gravimeter, 59, 63
 landfills, 32–33
 gravity methods in internal characterization of, 96–98, 97
 resistivity method in internal characterization of, 203–204
 latitude correction, 68
 layered media, seismic waves
 reflection and refraction of obliquely incident rays, 363–364
 reflection and transmission of normally incident seismic rays, 362–363, 364

leap-frog configuration, 216, 217
 Lee-partitioning electrode array, 175, 175
 Lesser Himalayas along the Purnea–Sevoke Transect, 95, 96
 line spacing, 483
 liquid scintillation, 401
 lithium reserves, geophysical exploration of, 18–19
 lithology logs
 gamma ray logging, 417–418
 self/spontaneous potential log, 418–419
 spectral logging, 418
 local earthquake tomography, 385
 locating archaeological sites in India, 453, 454
 logarithmic coordinates, 187
 logarithmic potential, 49
 logging while drilling (LWD), 420–421
 long-offset transient electromagnetic (LOTEM), 281, 285
 Loran-C, 429
 Lothal Revisitation Project, Gujarat, India, geospatial archaeology, 150–151
 Love wave, 360, 361
 low-pass filter, 135, *see also* high-pass filter

M

mafic-ultramafic contact, 437
 MAGMOD program, 16
 magnetic anomalies
 above Olivine Diabase Dyke, Pishabo Lake, Canada, 149, 149, **150**
 comparison of gravity with, 139–141
 maps, 153
 over magnetized sphere
 at the Equator, 141
 at the Northern Hemisphere, 141–143, 144
 at the North Pole, 141, 142
 when $i=22.5^\circ$, 144, 145
 when $i=67.5^\circ$, 143, 145
 over simple shapes, 139–144, 145
 magnetic cleanliness, 131
 magnetic data, 16
 magnetic dipoles, 117–118
 field lines for, 118
 magnetic effects, over dyke trending, 125, 125
 magnetic elements, 123
 magnetic equator, 122
 magnetic exploration, 115–154
 applications of magnetic surveying, **115**, 145–151
 case study, 145–151
 data interpretation, 138–139
 data processing, 134–138
 differences between gravity and magnetics, 116
 Earth's magnetic field, 123–124, 124
 measuring of, 127–130
 temporal variations of, 126–127
 field procedures, 130–133
 format for magnetic surveying, 133, 133–134
 geomagnetic data, 151–153
 magnetic anomalies over simple shapes, 139–144, 145
 magnetic elements, 123
 magnetic field nomenclature, 122–123
 magnetic monopoles, 116–119
 magnetics and geology, 124–126, 125
 magnetization of materials, 119

- mechanisms for induced magnetization, 119–120
- presentation of results (deliverables), 139
- remanent magnetization, 121–122
- similarities between gravity and magnetics, 115–116
- susceptibilities of rocks and minerals, 120–121, **121**
- magnetic field, Earth's, 122, 123–124, *124*
 - direction and component of, 130, *130*
 - diurnal variations, 126, 127
 - magnetic storms, 126, 127
 - measuring of, 127–130
 - magnetometers, 128–130
 - mechanical instruments, 127, *128*
 - secular variations, 126
 - temporal variations of, 126–127
- magnetic induced polarization (MIP), 239, *see also* induced polarization
 - survey over Teck's Highland Valley in British Columbia, Canada, 239
- magnetic method, 5–6
 - airborne (*see* airborne magnetic method)
- magnetic monopoles, 116–119, 504
 - forces associated with, 117
 - magnetic dipoles, 117–118
 - units associated with, 118–119
- magnetic observations, 130–131
- magnetic permeability, 540
- magnetic poles, 504
 - magnetic field, 122
- magnetic potential, 506
- magnetic properties
 - dipolar magnetic field, 505
 - diurnal variations, 507
 - field measurements, 507
 - field nomenclature, 506–507
 - instruments, 507
 - magnetic monopoles, 504
 - magnetic poles, 504
 - magnetic potential, 506
 - magnetic storms, 507
 - material magnetization, 505–506
 - Maxwell equation, 506
 - secular variations, 507
- magnetic resonance sounding (MRS), 32
- magnetics, **27**
 - and geology, 124–126, *125*
 - and gravity, differences between, 116
 - and gravity, similarities between, 115–116
- magnetic saturation, 153n11
- magnetic storms, 126, 127, 507
- magnetic surveying
 - applications of, **115**, 145–151
 - format for, 133, *133–134*
- magnetic susceptibility, 119, 540
- magnetization
 - induced, 119
 - of materials, 119
 - remanent, 119, 121–122
- magnetometers, 128–130
 - fluxgate magnetometer, 128–129
 - proton precession magnetometer, 129–130
- magnetotelluric (MT), 320–352
 - 3-D galvanic distortion, 333
 - applications of, 341–352
 - crustal research, 345–349
 - Earthquake precursor prediction research, 349–352
 - geothermal exploration, 344
 - hydrocarbon exploration, 343
 - mineral exploration, 343–344
 - assumptions of, 324
 - case study, 341–352
 - electromagnetic fields, 322–324
 - electromagnetic induction in homogeneous half-space, 324–327
 - field layout, 339–340
 - impedance tensor, 327–331
 - instruments used in, 339
 - inversion routines, 336–338
 - natural source low-frequency EM
 - Helmholtz wave equation, 522
 - Maxwell equations, 522
 - phase tensors, 526
 - resistivity and phase, 522–523
 - response functions, 523–525
 - rotation-invariant MT parameters, 525–526
 - skin depth, 522
 - static shift, 525
 - 3-D galvanic distortion and decomposition, 525
 - phase tensors, 335–336, *336*
 - processing and interpretation of data, 341
 - rotation-invariant magnetotelluric impedance tensors, 333–335
 - source of signal, 320–322
 - static shift, 333
 - strike analysis, 331–332
 - survey groups in India, **342**
 - variant of, 338–339
- main field, 123
 - variations, 132–133
- main magnetic field, 129–130
- man-made electrical noise, 247
- man-made radioactive isotopes, 438
- mapping soil, 249, 249–250
- The Marine and Coastal Survey Division (M&CSD), 471
- marine geophysical methods, **2**
- marine geophysics, *see also* geophysics; surveys and case histories, marine geophysics
 - dataset corrections, 482
 - deep-sea nodules exploration, blue economy, 484–485, **485**, *486*
 - exclusive economic zone (EEZ), 471
 - exploratory/research vessels, 483
 - geophysical methods, 526
 - gravity/magnetic survey, 478–481
 - integration of temporally different data, 483
 - interferometric sonar, 476–477
 - line spacing, 483
 - marine resistivity survey, 481–482, *482*
 - multibeam echo sounder (MBES), 487
 - multibeam sonar, 476
 - multichannel and 3-D seismic reflection surveying, 478
 - opportunities for India, 485–486, **487**
 - overview, 471–472
 - polymetallic nodule mining project, 485, *486*
 - propagation of seismic waves, 472–473
 - sediments, 473–474
 - side-scan sonar system, 474–475
 - single-beam echo sounders, 475–476
 - subbottom profiler survey, 477, 477–478

survey data processing, 483–484
 surveys and case histories, 487–492
 swath bathymetry survey, 475
 marine geotechnical and geophysical surveys, 38–39
 marine gravity survey, 64–65, 490–491
 marine magnetic survey, Israel and Libya, 491
 marine magnetometer survey, Tobruk, Libya, 492
 marine magnetotelluric (MMT), 338–339
 marine resistivity survey, 481–482, 482
 mass of the Earth, 48
 material magnetization, 505–506
 Maxwell equations, 506, 522
 Maxwell-Faraday equation, 263
 measured potentials magnitude, 248
 measurement errors, IP, 247–248
 measurement of SP/field procedure
 potential gradient method, 216, 217
 total field method, 217, 217–218
 mechanical instruments, 127, 128
 membrane polarization, 235, 235
 memory data, wireline logging, 411
 memory log, 421
 meter on board ship, 64–65
 meter onto the seafloor, lowering of, 64
 micro-anisotropic, 193
 mineral deposits
 gravity maps of, 105–106, 105–107, **107**, 108
 gravity signatures of, 109, **109–110**
 mineral exploration, 282–284, 343–344, 465
 in small-scale mining, Fazenda Santana, Brazil,
 256–258, **257**, 258
 sulphide ore body detection, 253
 mineral grains, 235, 235
 mineral mapping techniques, South-Eastern Rajasthan, 451
 Misti Volcano, Peru, 220–221
 Mixture Tuned Matched Filtering (MTMF)
 algorithms, 451
 mobile ERT imaging system, 197
 modelling gravity anomalies, 85–93
 of a horizontal cylinder, 89–90, 89–90
 horizontal line element, 87–89
 horizontal slab model for vertical fault, 92–93
 horizontal thin sheet, 90–92
 of point mass or sphere, 85–87, 86
 monostatic antenna, 287
 MT, *see* magnetotelluric
 mud logging, 422
 multibeam bathymetric surveys (MBBS), 487
 multibeam echo sounder (MBES), 487
 multibeam sonar, 476
 multichannel analyses of surface waves (MASW), **26**
 multichannel configuration, 340
 multiple-gradient array, 243
 multitransient electromagnetic (MTEM), 281
 Mundgod-Siruguppa, Dharwar Craton, South India,
 crustal research, 346
 Muragl
 geophysical imaging of Alpine Rock Glaciers, 304
 tomographic geoelectric survey, 202–203
 Murtèl
 geophysical imaging of Alpine Rock Glaciers,
 286–287, 303–304
 tomographic geoelectric survey, 202

N

Narmada-Son lineament (NSL), Madhya Pradesh (MP),
 94–95, 95
 crustal research, 345–346, 347–348
 earthquake monitoring, 349–351, 350
 National Aeronautics and Space Administration
 (NASA), 444
 National Geophysical Mapping Programme (NGPM), 44
 National Institution of Transforming India (NITI), 486
 natural currents, 213–227
 interpretation of SP data, 218
 self-potential (SP), 213–215, 215, **216**
 anomaly of sulphide ore body, Sariyer Area,
 Istanbul, Turkey, 224, **225**, 225
 application of, 218–219, 219–220
 to assess groundwater flow in hydrothermal
 systems, 220–224
 measurement of (*see* measurement of SP/field
 procedure)
 survey equipment, 215
 natural electric potentials, 540
 Navigation Satellite Timing and Ranging Global
 Positioning System (NAVSTAR GPS), 455
 negative anomalies, 105, *see also* anomaly
 negative self-potential anomaly, 219
 neutron detection, 400
 neutron porosity, 414–416, 415
 Newtonian potential, 49
 nickel sulphide exploration, Kambalda District, Western
 Australia, 147–148, 422–423
 non-linear conjugate gradient (NLCG), 337
 non-uniqueness of solution, 85
 2-D gravity modelling, 85
 Northridge, California earthquake, coseismic
 displacements, 458
 nuclear magnetic resonance (NMR) logging, 420

O

OCCAM 2-D MT software, 434–435
 Occam inversion, 337
 Ohm's law, 159–160, 160, 513–514
 oil and gas exploration, Ethiopia, 464
 optically pumped magnetometer, 481
 ore bodies, 109
 ore compositions, 109
 orientation of lines, 531
 The Outer Thames Estuary Regional Environmental
 Characterisation (REC), 489
 overhauser magnetometers, 480

P

Palghar Region, Maharashtra, earthquake
 monitoring, 351
 paramagnetism, 119–120
 paths in layered media, 509–510
 pendulum method, absolute gravity measurement on Earth
 surface, 57–58
 periodic repeat reading, 532
 permittivity constant, 156
 Peruvian volcanoes, 221

- petrophysical/geophysical parameters, spatial distribution of, 28
- phase domain, 237–238
- phase tensors, 526
 - magnetotelluric (MT), 335–336, 336
- photomultiplier tube (PMT), 400
- plane interface
 - distortion of current flow at, 180, 180
 - distortion of potential at, 181–182, 181–184
- plane wave source of excitation, 273–278
 - very-low-frequency (VLF), 273–274, 273–276
 - VLF resistivity and radio magnetotelluric (RMT, CSRMT) method, 276–277, 276–278, **278**
- plate motion models, 459–460
- point mass
 - modelling gravity anomalies, 85–87, 86
 - schematic diagram of, 62
- polarization, of the Earth, 233–237
- polar satellite launch vehicle (PSLV), 446
- pole-dipole array, 176, 176, 242
- pole-pole array, 176–177, 177, 243
- polymetallic nodule mining project, 485, 486
- porosity logs
 - density logging, 413–414
 - neutron porosity, 414–416, 415
 - sonic log, 416–417
- porosity of rocks, **165**
- positive anomalies, 105
- possible triassic folding below Abu Rassain, the Western Desert, Iraq, 100–102
- potential(s), *see under* principles
 - amplitude, 217, 217–218
 - decay of, 234
 - electric field and, 158–159, 159
 - electrochemical, 218
 - field equations, 49–50
 - gradient method, 216, 217
 - in homogeneous media, 162–163
 - three-dimensional, 49
- potential fields, 43–110
 - gravity anomalies, 71–72
 - Bouguer anomaly, 72
 - data format for reduction of gravity data, 72, **72**
 - free air anomaly, 71
 - isostatic anomaly, 72
 - gravity data of India for delineation of structural fabric and crustal architecture, 103–104, 104
 - gravity maps of mineral deposits, 105–106, 105–107, **107, 108**
 - gravity signatures of mineral deposits, 109, **109–110**
 - gravity surveying, 44, 65–71
 - applications and case study, 93–103
 - reduction of gravity data, 67–71
 - interpretation of gravity anomalies, 77–93
- isostasy
 - Airy-Heiskanen model, 73, 74
 - isostatic compensation and vertical crustal movements, 74–75
 - Pratt-Hayford isostatic model, 73–74, 74
 - Kepler's Laws of Planetary Motion, 44–46, 45–46
- measurement of gravity on land
 - in boreholes, 63–64, 64
 - on the Earth's surface, 57–63
- measurement of gravity on moving platforms
 - air surveys, 65
 - sea surveys, 64–65
 - space measurements, 65
- principles, 46–57
 - geoid of reference, 54, 55
 - gradients of gravity, 51
 - gravitational potential, 48–50
 - horizontal gradient of gravity from top and bottom of tower, 51
 - mass of the Earth, 48
 - second derivative, 51
 - shape of the Earth, 52, 53
 - spheroid of reference, 54, 55
 - units of gravity, 47
 - variation of g from place to place, 55–57, 56
 - vertical gradient of gravity from top and bottom of tower, 51
- rock density, 75–77, **76, 77**
- Pratt-Hayford isostatic model, 73–74, 74
- precession, 153–154n12
- precision, 531
- principles
 - electrical, 157–164
 - atomic charge, 161
 - electrical conduction, types of, 161–162, **162, 162**
 - electric field and potential, 158–159, 159
 - equipotentials, 161
 - Ohm's law, 159–160, 160
 - potentials in homogeneous media, 162–163
 - single-current electrode at depth, 163–164, 164
 - of electromagnetic (EM), 263–264
 - of equivalence, 192
 - of induced polarization (IP), 233–237
- potential fields, 46–57
 - geoid of reference, 54, 55
 - gradients of gravity, 51
 - gravitational potential, 48–50
 - horizontal gradient of gravity from top and bottom of tower, 51
 - mass of the Earth, 48
 - second derivative, 51
 - shape of the Earth, 52, 53
 - spheroid of reference, 54, 55
 - units of gravity, 47
 - variation of g from place to place, 55–57, 56
 - vertical gradient of gravity from top and bottom of tower, 51
- of resistivity method, 169–174, 170–173, 514
- probing depth, 187, 516
- propagation
 - electromagnetic (EM) induction, 520
 - electromagnetic wave, 289–290
 - of seismic waves, 472–473
- proton precession magnetometers, 129–130, 480
- proton-rich nuclei, 397
- pseudorange measurements, 428
- pseudosection, 194, 195, 205, 244, 245, 250, 254, 350
- PTHEM, Aladahalli Area, Hassan District, Karnataka, 452, 453
- pulse generators, 295, 296, 297
- pure ferromagnetism, 120

Q

quadrature component, 266–269, 519

R

radioactive dating, 407

radioactive method, 8

radioactivity

alpha decay, 394, 394–395, 397

alpha radiation, 398

atoms, 391–392

beta-minus decay, 395

beta-plus decay, 396, 397

beta radiation, 398

earthquake precursors, **405**, 405–407, 406

electron capture, 396–397, 397

emissions, 393–394

gamma decay, 395, 395–396, 397

gamma radiation, 398

Geiger-Muller (GM) counter

operating principle, 399–400

types and applications, 400

in geophysical exploration

buried channel and flow of underground fluids, 403–405, 404

minerals and metals, 402–403

surface gamma activity, 403

half-life, 392–393, **393**

isomeric transition, 396

overview, 391

radioelements and detectability, 402

scintillation counter

definition, 400

effective use of, 401–402

efficiencies of, 401

operational mechanism, 401

photomultiplier tube (PMT), 401

units of, 399

radioactivity decay, 392, 398

radioactivity decay series, 397, 398

radioelements and detectability, 402

radio magnetotelluric (RMT), 262

radiometric assaying, 440, 440

radiometric method, airborne, *see* airborne radiometric method

radiometric resolution, 449

rapid deformation, South Flank of Kilauea volcano, Hawaii, 460

ray geometry and travel times, 293–294

Rayleigh wave, 360, 361

real-time data, wireline logging, 411

receiver systems, 296–297

reciprocity, 174

property of, 174

reduced-to-the-equator (RTE), 138

reduction-to-pole (RTP), 137, 138

reflection and refraction

electromagnetic (EM) induction, 520–521

of plane waves, 291, **292**

seismic method, 6

seismology, 380

surveying, 510–512

tomography, 385

refraction microtremor (ReMi), **26**

regional basement modelling, 12–18, 15

dipole-dipole sections, 17, 17–18

gradient IP map, 16

gradient resistivity map, 16

magnetic data, 16

self potential (SP), 16

regional geology, 33–34

regional/global tomography, 385

regional-residual gravity anomaly, 77–78, 78

depth of anomalous mass, estimation of, 107

separation of, 107, 108

regional-residual separation, 134–135

relative gravity, measurement of, 59–61, 60

sensitivity of unstable or astatized types of gravimeter, 59–61, 60

remanent magnetization, 119, 121–122, 540

remote reference magnetotelluric (RRMT), 340

remote sensing, 9

case study, 451–454

in current usage, 444

data acquisition techniques, 446

data processing, 449–450

electromagnetic radiation (EMR), 444

hyperspectral vs. multispectral remote sensing, 450

overview, 445

radiometric resolution, 449

satellite and data overview, **446–448**

software, 450

spatial resolution, 449

spectral resolution, 449

temporal resolution, 449

thrust areas in applications, 449

resistivity, 161, 175, 514

for archaeological site characterization of Jordan, 204

highest resistivity, 166

imaging, 34

intermediate resistivity, 166

in internal characterization of landfills, Ilhavo, Portugal, 203–204

log, 412

lowest resistivity, 166

measurement, 177, 179

and phase, 522–523

of rocks, soils, and ores, 167

surveying, 185–186, 516–518

response functions, 523–525

RINEX observation file, 457

Risby landfill, field sites and data collection, 33

rise-and-fall method, absolute gravity measurement on Earth surface, 59

rock density, 75–77, **76**, 77

rock resistivities, 165, **166**, 167

rocks and minerals, susceptibilities of, 120–121, **121**

Ronka, *see* horizontal loop EM (HLEM)

rotation-invariant MT parameters, 525–526

RV *Samudra Manthan*, 471

S

saltwater intrusion, 285–286

sampling, 9–10

airborne geophysics, 426

Sanriku-Haruka-Oki earthquake, Japan, 459

- Saurashtra Region, Gujarat, India, hydrocarbon exploration, 343
- scattering and diffraction, 291–292
- Schlumberger array, 175, 175–176, 186, 242
- Schlumberger electrical resistivity data sheet, 534
- scintillation counter
 - definition, 400
 - effective use of, 401–402
 - efficiencies of, 401
 - operational mechanism, 401
 - photomultiplier tube (PMT), 401
- sea surveys, measurement of gravity, *see* marine gravity survey
- second derivative, 51, 79
- secular variations, 126, 507
- sediments, 473–474
- seismic/earthquake studies, Aizawl Town, Mizoram, 466
- seismic microzonation, Patna town by GSI, 388
- seismic reflection method
 - field survey, 375–378
 - single dipping reflector, 371–375, 374, 375
 - single horizontal reflector, 369–371, 370, 372
 - two-way travel time (TWTT), 368
 - vertical and horizontal resolution, 378
 - vertical reflected ray paths, 368
- seismic refraction software, 465
- seismic refraction surveying
 - geographic information system (GIS), 465
 - horizontal interface, 379, 379–381
 - inclined interface, 381, 381–383
- seismic sources
 - airguns, 366–367
 - characteristics, 365
 - explosives, 366
 - hammer, 365
 - seismic vibrator, 367
 - seismograph/seismometer, 367
 - transducers, 367
 - weight-dropping thumper technique, 365–366
- seismic surveying, 358
- seismic tomography
 - applications of, 385–386
 - defined, 383
 - history, 384
 - limitations, 386
 - local tomographic models, 384–385
 - process, 384–386
 - regional/global tomography, 385
 - theory, 383–384
- seismic vibrator, 367
- seismic waves
 - axial modulus, 359
 - body waves, 359–360, 360
 - compressional waves (*P* waves), 359, 360
 - shear waves (*S* waves), 359, 360
 - bulk modulus, 359
 - earthquake magnitude, 512–513
 - layered media
 - reflection and refraction of obliquely incident rays, 363–364
 - reflection and transmission of normally incident seismic rays, 362–363, 364
 - paths in layered media, 509–510
 - propagation of, 472–473
 - reflection surveying, 510–511
 - refraction surveying, 511–512
 - shear modulus, 359
 - stress and strain, 358–359, 507–508
 - surface waves
 - Love wave, 360, 361
 - Rayleigh wave, 360, 361
 - types of, 508
 - velocities, 509
 - and densities, **362**
 - elastic wave velocity, 361
 - time average equation, 361–362
 - Young modulus, 359
- seismogram, 365
- seismograph/seismometer, 367
- seismological data, India
 - Assam Valley by GSI, 387
 - earthquake studies, 387
 - Patna town by GSI, 388
 - Shillong Plateau, 387
 - Tripura fold belt, 387
- self-potential (SP), 8, 16, **27**
 - anomalies, 16, 214, 223
 - data sheet, 537
 - defined, 215
 - noise, 248
 - in oxidation (redox) potential, 215
 - sulphide ore body, anomaly of, 224
 - as tool to assess groundwater flow, 220–224
 - transient anomaly, 223–224
- self/spontaneous potential log, 418–419
- sensitivity, 531
- spherics, 247
- shallow geophysical investigation, 28
- shallow seismic reflection, **26**
- shallow seismic refraction, **26**
- shape of the Earth, 52, 53
- shear modulus, 359
- shear waves (*S* waves), 359, 360
- shipborne, magnetic observations, 130
- side-scan sonar system, 474–475
- signal travel times, EM induction, 521
- Siliguri-Gangtok, Sikkim-Darjeeling Himalaya, crustal research, 345
- Silurian Reef exploration, 490–491
- single-beam echo sounders, 475–476
- single-current electrode at depth, 163–164, 164
- single dipping reflector, 371–375, 374, 375
- single horizontal reflector, 369–371, 370, 372
- site geometry, 439–440
- skew, 331
- skin depth, 264–265, 327, 522
- Slingram, *see* horizontal loop EM (HLEM)
- Sohagpur, Central India, magnetic surveying, 146, 146
- sonic logging, 416–417
- Southern Granulite Terrain (SGT), crustal research, 348–349
- space measurements, 65
- spatial distribution, petrophysical/geophysical parameters, 28
- spatially varying corrections, 132
- spatial resolution, 449
- Spectral Angle Mapper (SAM), 451
- spectral induced polarization (SIP), 237

spectral logging, 418
 spectral noise logging (SNL), 420
 spectral resolution, 449
 spheroid of reference, 54, 55
 split-spread method, 375, 376
 spontaneous potential, *see* self-potential (SP)
 Stanthorpe, residual anomaly map of, 15
 static shift, 525
 effects of, 333
 station spacing, 531
 statistics, airborne radiometric, 439
 storm monitor, 532
 stress and strain, seismic waves, 358–359, 507–508
 strike analysis, magnetotelluric (MT), 331–332
 Stromboli volcano, 221
 structural fabric, 103–104
 structural index, 81, **81**
 subbottom profiler survey, 477, 477–478
 subduction zones, 386
 subsurface voids, 200
 sulphide ore body
 detection, 253
 Sariyer Area, Istanbul, Turkey, 224, **225**, 225
 surface fitting, 78
 surface geophysical methods, **2**
 surface waves
 Love wave, 360, 361
 Rayleigh wave, 360, 361
 survey data processing, 483–484
 surveys and case histories, marine geophysics
 baseline geophysical data for hazard management,
 coastal areas, 488–489
 buried volcanic structures, Gulf of Naples,
 491–492, 492
 continental margin of North-West Ireland, 489–490
 Curie isotherm depth (CID), 487
 German Antarctic North Victoria Land Expedition
 (GANOVEX) expeditions, 490
 marine gravity survey, 490–491
 marine magnetic survey, Israel and Libya, 491
 multibeam bathymetric surveys (MBBS), 487
 The Outer Thames Estuary Regional Environmental
 Characterisation (REC), 489
 Torre del Greco volcanic structure, Naples Bay,
 Southern Tyrrhenian Sea, Italy, 489
 unexploded ordnance, Tobruk, Libya, 491
 swath bathymetry survey, 475
 syncline, 89, 89–90

T

Teck's Highland Valley in British Columbia, Canada, 239
 teleseismic tomography, 385
 telluric currents, 224–227, 227, *see also* natural currents
 telluric noise, 247
 temporal resolution, 449
 temporal variations, 131–132
 TEM/TDEM techniques, *see* transient electromagnetic
 terrain correction, 70–71
 terrigenous sediments, 473
 thermal water, Shear Fault (Colorado), 222
 thermoremanent magnetization (TRM), 121
 three-dimensional potential, 49
 tidal correction, 68

time average equation, 361–362
 time domain array sounding (TDAS), 284
 tipper, 331–332
 tomographic geoelectric survey
 Muragl, 202–203
 Murtèl, 202
 Tons Valley base metal prospect, Uttar Pradesh (UP),
 465–466
 Torre del Greco volcanic structure, Naples Bay, Southern
 Tyrrhenian Sea, Italy, 489
 total field measurements, 129–130
 total field method (fixed-base) configuration, 217, 217–218
 transducers, 367
 transient electromagnetic (TEM/TDEM) techniques,
 279–287
 geophysical imaging of Alpine Rock Glaciers, Murtèl,
 286–287
 hydrocarbon exploration, 284–285
 mineral exploration, 282–284
 saltwater intrusion with, 285–286
 transient self-potential anomaly, 223–224
 TRANSIT Navy Navigation Satellite System
 (NNSS), 455
 travel time vs distance, 370, 373, 379
 traverse/line spacing, 531
 trend surface analysis, 135
 Tuwa-Godhra Profile, Gujarat, crustal research, 347
 Twin Otter Airborne Survey System (TOASS), 444
 two-way travel time (TWT), 368

U

ultramafic-mafic contacts, 422
 unaccountable error, 532
 underwater ERT survey, 197
 unexploded ordnance, Tobruk, Libya, 491
 units
 of gravity, 47
 magnetic monopoles, 118–119
 of radioactivity, 399
 upward continuation, 135, 136
 Uranium-238, 398
 UTEM (the University of Toronto EM) system,
 280–281

V

vadose zone, 30–31, 30–32
 variation
 of g from place to place, 55–57, 56
 of g with altitude and depth, 56
 velocities, 509
 and densities, **362**
 elastic wave velocity, 361
 time average equation, 361–362
 vertical resolution, 378
 vertical contact, signal enhancement in case of, 80
 vertical crustal movements, 74–75
 vertical derivative, 136
 vertical electrical sounding (VES), 189–194
 ambiguities of sounding curve interpretation,
 192–194
 in locating fresh water at VizagSteel Plant,
 Visakhapatnam, Andhra Pradesh, 204–205

- over horizontally stratified media, 189–192, *191*
 - over laterally inhomogeneous media, 192
- vertical fault
 - gravity anomaly across, 93
 - horizontal slab model for, 92–93
- vertical gradient of gravity from top and bottom of tower, 51
- vertical reflected ray paths, 368
- very-low-frequency (VLF), 273–274, 273–276
 - resistivity and radio magnetotelluric (RMT, CSRMT) method, 276–277, 276–278, **278**
- VES, *see* vertical electrical sounding
- viscous remanent magnetization (VRM), 121
- volcano
 - buried volcanic structures, Gulf of Naples, Italy, 491–492, *492*
 - Misti Volcano, Peru, 220–221
 - Miyakejima, 220–221
 - Mt Etna, 460–462
 - Mt Fuji, 220–221
 - Torre del Greco volcanic structure, Naples Bay, Southern Tyrrhenian Sea, Italy, 489
- Vulcano (Italy), groundwater flow pattern, 222

W

- Wagad Region, Gujarat, earthquake monitoring, 351–352
- wave equation tomography, 384–385
- wavelength filtering, 135
- wavelength of anomalies, 105
- weight-dropping thumper technique, 365–366
- Wenner array, *174*, 174–175, 186, 241, 242
- Wenner electrical resistivity data sheet, 533
- Western Desert, Iraq, triassic folding, Abu Rassain, 100–102
- Widgie Townsite nickel prospect, Western Australia, 435
- wireline logging
 - history of, 411–412
 - memory data, 411
 - real-time data, 411

Y

- Young modulus, 359

Z

- zagros lithosphere, Kermanshah, Iraq, earthquake, 451