CHAPTER FIFTEEN

Geophysical Imaging of the Continental Lithosphere

An essay by Frederick A. Cook

15.1 INTRODUCTION

The lithosphere beneath the continents is a vast and largely unexplored region of the Earth. Because it is inaccessible to normal geologic observations, various geophysical tools, including measurements of magnetism, gravity, electricity, subsurface temperatures, and earthquake waves, have been used to gather information about it for more than 100 years. Since about 1975, however, application of controlled-source **seismic reflection methods** has produced images of the deep subsurface that are visually similar to geologic cross sections and are therefore readily accessible to most earth scientists. When analyzed in conjunction with other **geophysical imaging methods,** as well as with the known geology, these data provide the most detailed information on subsurface structure presently available.

In many areas of the continents, preliminary reconnaissance has been accomplished and a few major features have been discovered: some of these are extensions of the surface geologic context; others resemble features we know but cannot be directly linked to our geologic base; still others are unusual and outside our frame of reference. Overall, we have only begun the search.

15.2 WHAT IS SEISMIC IMAGING?

Seismic imaging methods use vibrational (elastic) energy generated on or near the Earth's surface as a source of waves that propagate into the subsurface, reflect from or refract through interfaces at depth, and then return to the surface where they are digitally recorded for subsequent data enhancement. Refracted waves are valuable to delineate regional characteristics of the crust and upper mantle, including the base of the crust, whereas reflected waves are useful for mapping structural detail. The source of energy may be produced artificially, as with an explosion or vibratory signal, or it may be natural, as with an earthquake. In general, artificial sources do not have as much energy as earthquakes do, and hence do not penetrate as deeply, but the results derived from artificial sources have much finer detail and are easier to relate to known geologic features. Seismic reflection profiling is not new. It was first employed by the petroleum industry about 75 years ago in oil exploration. However, many refinements, particularly following the advent of digital technology in the 1960s and 1970s, have made it possible to obtain images from greater depths and with greater precision than was considered feasible even just a few years ago.

The **seismic reflection method** is conceptually simple but logistically intensive (Figure 15.1). In most applications today, the method includes four essential components:

1. A vibrational energy source, usually several (3–5) synchronized, truck-mounted vibrators (to +/– 0.001 s or less). The elastic energy radiates into the subsurface, reflects off boundaries at depth, and returns to the Earth's surface where the resulting ground motion is measured by a series of miniseismometers (geophones). Once a vibration point is completed, the vibrator trucks move a few tens of meters to the next point and repeat the process.

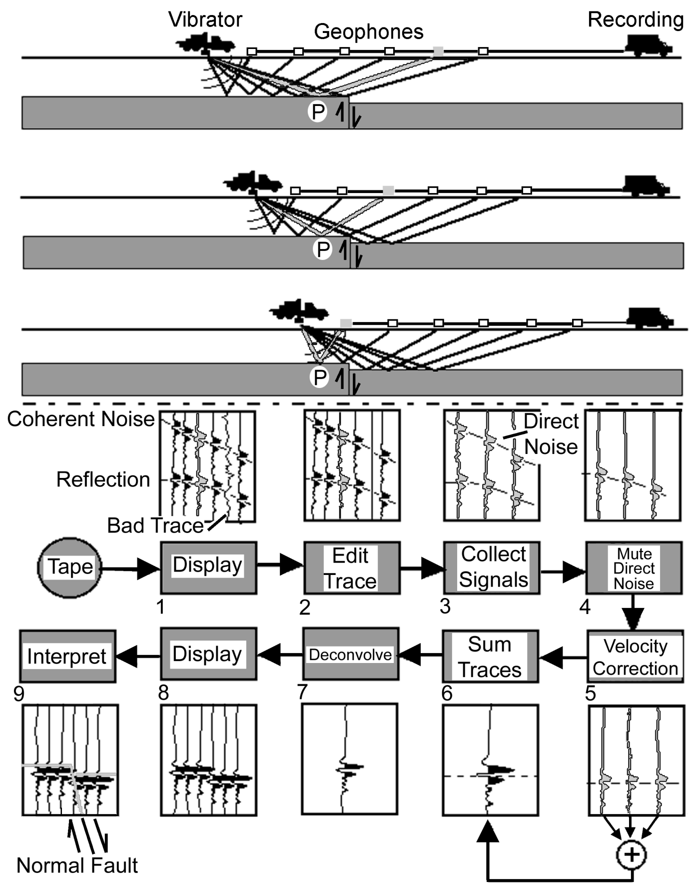

FIGURE 15.1 Schematic diagram illustrating the principles of seismic reflection profiling. At the top, three steps in field acquisition are shown to indicate how a reflection from a single subsurface point (P) is recorded by several different positions of the vibrator sources and the geophone receivers (gray geophone). This process, known as a common midpoint, or CMP, acquisition allows the recorded traces from P to be analyzed in the data processing steps (below) for the purposes of improving the signal.

2. A line of geophones. Each geophone is typically slightly larger than a 35-mm film container, and there are usually several thousand geophones spread over a single line to receive the signals from a single vibration location. The line of receivers is commonly several kilometers long (10–12 km or longer for most modern lithosphere-scale reflection surveys). As the sources (e.g., vibrator trucks) move for each source point, the receiving line also moves. For a survey that is 500 km long, there may be 10,000 vibration points.
3. A recording system, usually a box with computers and recording media (e.g., digital tapes) mounted on a truck. The signals are transmitted to the recording truck where they are stored for later computer enhancement. The electronics in the recording truck also provide the radio control that is sent to the vibrator trucks to initiate the vibrator signal.
4. A data processing system, which includes a suite of computer software that is applied in a series of steps to enhance the desired signal at the expense of unwanted noise. Some testing can be accomplished in the field to examine the data, but most of the intensive data processing requires weeks or months of effort to optimize the results.

The final product is a two-dimensional cross section of the earth that is displayed in terms of the signal transit time (the time required for a signal to follow a path from the surface source to a reflecting boundary at depth, and then to return to the surface). The display of the image in terms of transit time is the primary difference between a seismic cross section and a geologic cross section, as the "time section" needs to be converted to depth for an effective comparison to geologic boundaries. Once this is accomplished, the most fundamental result of all seismic profiles is an outline of the subsurface geometric framework.

Conversion to depth requires knowledge of the **seismic wave velocities** in the rocks. Although average velocities can be estimated for a general understanding of a cross section (we know that seismic waves generally travel faster in crystalline rocks than in sedimentary rocks), the more accurately the variations in wave velocity are known, the greater will be the accuracy of the image. Nevertheless, a good rule-of-thumb estimate for average seismic velocities in crystalline rocks of the crust is about 6.0 km/s; hence the depth in kilometers may be estimated by multiplying one-half of the transit time by a factor of 6.0. In this conversion a transit time of 5.0 s corresponds to a depth of about 15.0 km. Below the Mohorovičić discontinuity, the seismic wave velocity increases to about 8.0 km/s.

15.3 HOW ARE DATA INTERPRETED?

Great strides have been made in using seismic reflection profiling for mapping the internal structure of the lithosphere. Reflections from the deep crust (to a depth of about 30–40 km) were considered curiosities until the 1970s; reflections from the subcrustal lithosphere to twice this depth, or even more, are not unusual now. However, with technological advancements and the resulting improvements in signal quality and image depth, the challenges to interpretations have been great, because, as we obtain images from structures that are far removed from our geologic reference on the surface, the ability to relate them to what we know is increasingly limited. In many cases, ancillary geophysical methods that respond to different physical properties may be helpful in limiting possible interpretations, because the seismic images provide a geometric framework of structures that are found at many scales in the lithosphere.

Interpretation of the resultant seismic sections is in many ways analogous to interpreting a geologic cross section; indeed, the goal of the data processing is to provide an image that closely approximates a geologic cross section. However, careful interpretations require extensive knowledge of seismic wave propagation in rocks, geologic principles, and the regional geologic and geophysical context. Interpretation is often an iterative process: new images of subsurface geometry from the reflection profiles commonly spawn new geologic (or other geophysical) projects to test some of the ideas. Data resulting from these projects can then be used to review and often reinterpret the subsurface images. It is not unusual to rework data that may be 10 or 15 years old as new data processing techniques and new geologic information become available.

Although this essay is not intended to provide a complete review of image types or all major discoveries, it is helpful to consider images of a variety of common geologic features.

15.4 SOME EXAMPLES

Many of the criteria that are commonly used to identify faults in layered sedimentary rocks are not applicable in most crystalline rocks. For example, in layered strata, faults are usually delineated by offset layers, whereas reflections from a fault plane are rare. In generally unlayered crystalline rocks, however, offsets are difficult to observe (because it is not easy to correlate from one side of a fault to the other) and reflections from fault planes or fault zones are common.

In Figure 15.2a, for example, prominent layered reflections from Precambrian sills outline an anticline, the right (east) side of which is faulted by a west-dipping normal fault, the Rocky Mountain Trench in southwestern Canada. In this case prominent layering can be correlated across the fault to provide some estimate of the type (listric normal) and amount of displacement (about 10 km) along the fault. On the other hand, in Figure 15.2b reflections are visible from the Wind River fault zone in western Wyoming. Near the surface (to about 4 s travel time, or about 12 km depth), Precambrian crystalline rocks on the east are thrust westward over sedimentary strata of the Green River Basin on the west. Accordingly, the seismic velocity contrast between the crystalline rocks and the sediments is quite large, and a prominent reflection from the boundary is produced. The Wind River Thrust can then be followed as a series of subparallel reflections that downdip eastward to more than 7.0 s travel time (about 21 km depth).

Here then is a dilemma. If the contrast between the sedimentary rocks of the Green River Basin and the overlying crystalline rocks of the Wind River uplift is what produces the reflection along the shallow portion of the fault, then what is the cause of the reflection at greater depths where crystalline rocks are juxtaposed with crystalline rocks? Although this is discussed later in more detail, the answer in this case appears to be that the reflections are cause by mylonitic rocks that were formed as a result of the faulting. The process of **mylonitization** causes a very strong preferred orientation of crystals that, in turn, produces a contrast with the more randomly oriented crystals above and below. Thus, even where crystalline rocks are seismically homogeneous, the faulting process may produce surfaces that appear as subparallel reflections, particularly if faulting occurs below the brittle–plastic transition.

Seismic reflections from relatively homogeneous igneous rocks (e.g., plutons, basalt flows) are generally not easily observed. Exceptions are rocks that are deformed so that reflections may arise from the deformation surfaces (as described above for the Wind River uplift), and igneous intrusions (e.g., sills) that are sufficiently thin that there is a measurable contrast with surrounding rocks such as sediments or other crystalline rocks (Figure 15.2c).

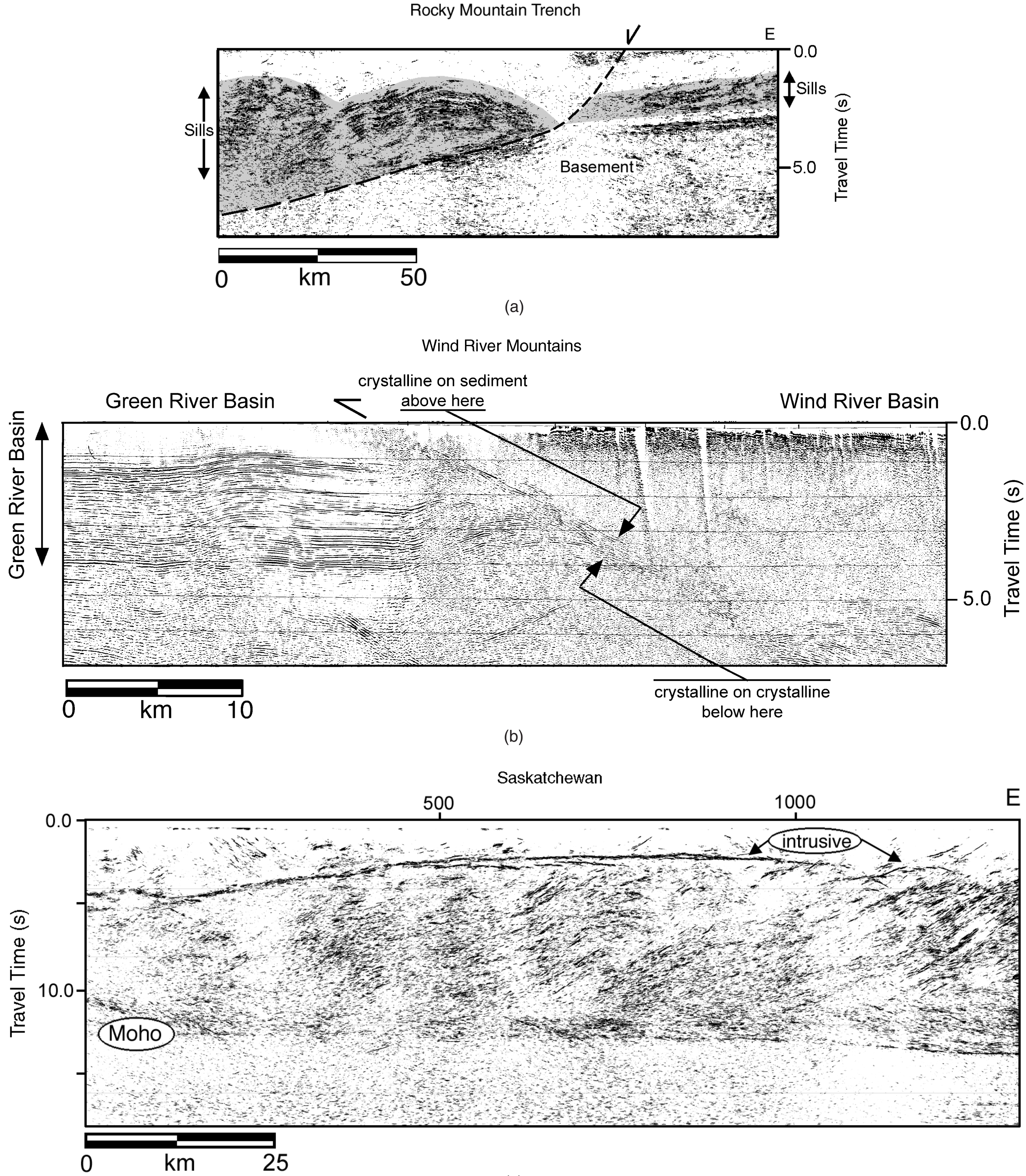

FIGURE 15.2 (a) Seismic reflection profile across the southern Rocky Mountain Trench near the Canada-U.S. border. Note that the prominent layering, which is drilled on the west and is known to be dominantly Proterozoic sills, is offset along a west-dipping listric normal fault that has about 10 km of dip-slip displacement. Data were recorded by Duncan Energy of Denver, Colorado. (b) Seismic profile from the Wind River Mountains in Wyoming (USA). The Wind River fault juxtaposes crystalline rocks of the Wind River Mountains with sedimentary rocks of the Green River Basin along a moderately east-dipping fault, and this provides a simple explanation for the prominent reflection. Below a travel time of about 3.5–4.0 s, however, the fault zone places crystalline rocks onto crystalline rocks and the reflections must be caused by other mechanisms. Data recorded by COCORP (Consortium for Continental Reflection Profiling) in 1977. (c) Seismic profile from the Proterozoic Trans-Hudson Orogen in northern Saskatchewan (Canada) illustrating prominent subhorizontal reflections that have been interpreted as intrusive rocks. Note that the reflector appears to cross cut several dipping reflections. Note also the prominent Moho on these data.

15.5 THE CRUST–MANTLE TRANSITION

The transition from the crust to the mantle is generally considered to be a relatively simple surface that has mafic rocks such as gabbro or mafic granulites above, and ultramafic rocks below (see Chapter 14). Indeed, much research in the 1950s to 1970s attempted to address the question of whether the transition is a compositional change (i.e., gabbro or granulite in the crust to peridotite in the mantle) or whether it could be a change in phase (as from mafic granulites in the crust to eclogites in the mantle). Central to the discussion was the observation from regional seismic refraction

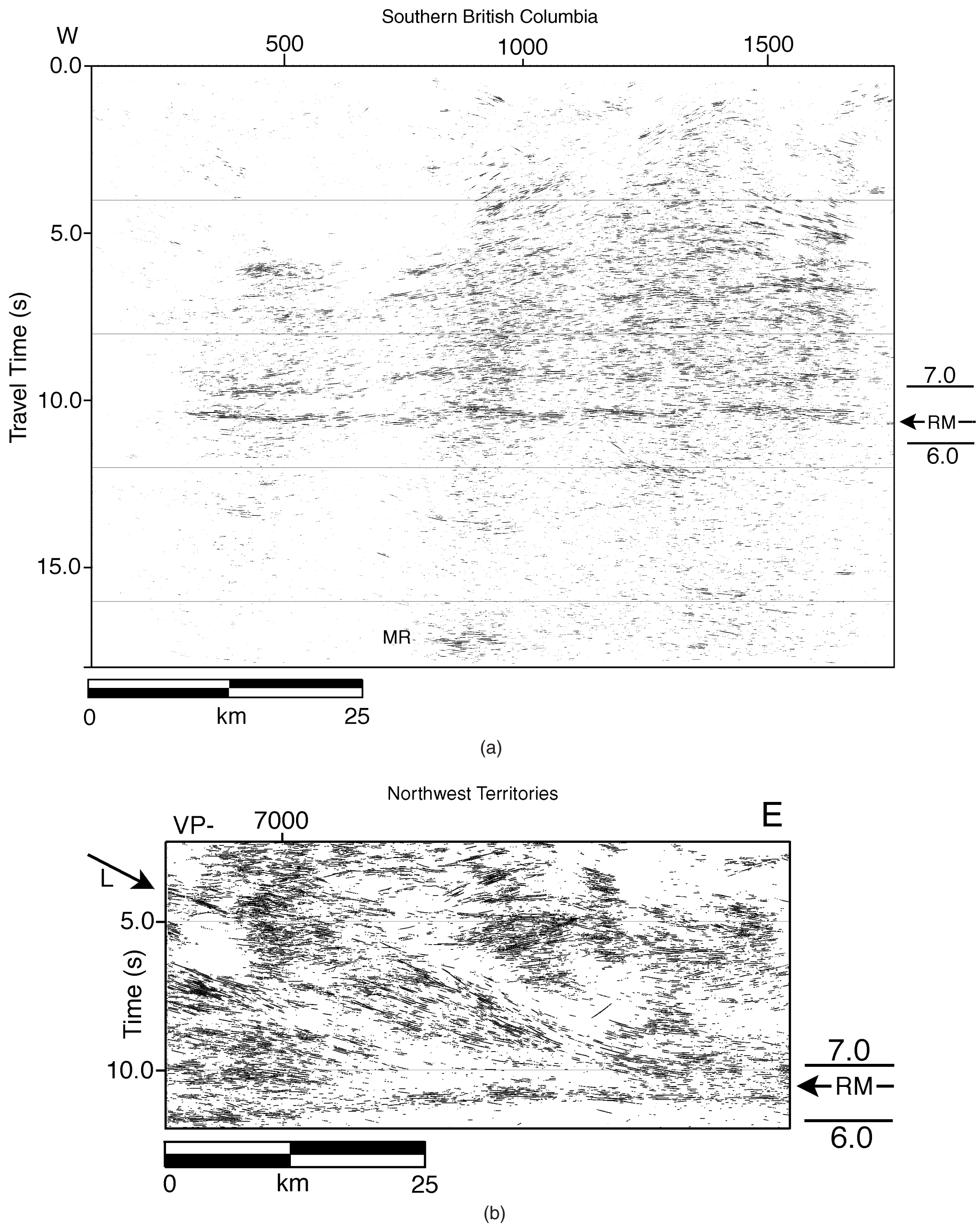

FIGURE 15.3 Some reflection characteristics of the crust–mantle transition. (a) Profile from south-central portion of the Canadian Cordillera illustrating a relatively simple, single reflection from near the transition. On the right side of the figure, the numbers 6.0 and 7.0 represent the positions of the Moho, as identified from adjacent seismic refraction data, for average velocities of 6.0 and 7.0 km/s, respectively. RM represents the preferred position of the Moho using the crustal velocity structure determined from the refraction profile.

and earthquake data that there is almost always a prominent seismic velocity increase at 40–50 km beneath the continents and about 10 km beneath the oceans. On these grounds, the boundary appears to be relatively simple and globally significant, and, as such, is known as the **Mohorovičić discontinuity,** or **Moho.**

As the information obtained from reflection profiling becomes increasingly detailed, however, we see that the crust–mantle transition is clearly not a uniform boundary because lateral variations in its geometry and reflection characteristics are common (Figure 15.3). It can be structurally complex or simple, multilayered or single

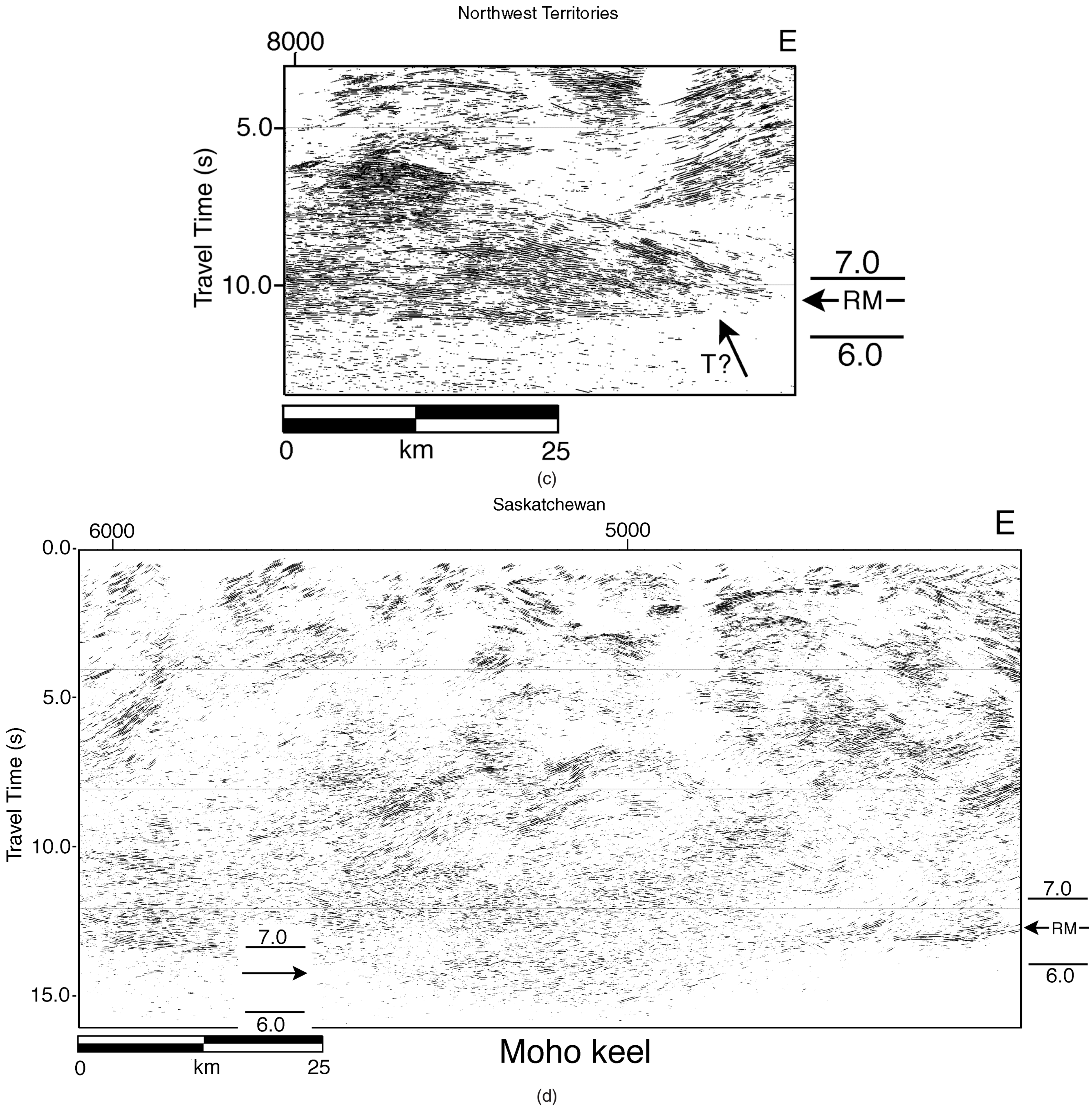

FIGURE 15.3 (Continued) Note that the Moho appears to be located at the base of crustal reflectivity, and that the underlying mantle has fewer reflections (e.g., MR). Data were recorded by LITHOPROBE in 1988. (b) Portion of a seismic profile that illustrates listric structures into the crust–mantle transition. Data were recorded by LITHOPROBE in 1996. This segment is from beneath the Great Bear arc region on the regional profile (Figure 15.5). (c) Portion of a seismic profile that illustrates many lower crustal layers that are parallel to the Moho as well as a possible truncation (T?). Data were recorded by LITHOPROBE in 1996. (d) Portion of a seismic profile from northern Saskatchewan (Canada) that illustrates a local deepening of the crust–mantle transition (Moho keel). Note that although there is not a prominent reflection near the transition, the reflectivity does diminish near it. In this figure, two locations for estimates of travel time to the reflection Moho are indicated from adjacent refraction profiles.

surface, flat or dipping; and any of these variations may be present along a single profile, sometimes changing over distances of only a few kilometers.

On the other hand, one of the most obvious characteristics of many reflection profiles is the transition from reflective crust to relatively nonreflective mantle (Figure 15.3). Indeed, this effect is so pervasive on a global scale that it is commonly used as a means to identify the crust–mantle transition; the "reflection Moho" is generally interpreted to be at the base of prominent crustal reflectivity. Thus, on one hand the detailed structural and reflection characteristics of the transition are complex and variable (Figure 15.3), while on the other there are regional, large-scale differences in the reflectivity of the crust and the upper mantle. Until the advent of crustal reflection profiling, and particularly high-quality detailed images of the lower crust and upper mantle, these characteristics were not observable. As a result, any future interpretation of the crust–mantle transition must account for geometric complexities at relatively small scales (kilometers to tens of kilometers), and relative uniformity when viewed at larger scale with lower resolution. This may ultimately be one of the most fundamental results of these kinds of data as it will lead to new concepts of how the crust and mantle interact.

One of the more heated debates in the interpretation of deep-crustal reflection profiles has been the cause (or causes) of reflectivity. Some aspects are well understood. For example, a reflection must result from a change in seismic velocity and/or rock density and the magnitude of the reflection (amplitude) is related to the magnitude of the contrast. Hence, a contrast between a rock with relatively low seismic velocity, such as a sandstone, and another with a relatively high seismic velocity, such as a gneiss, will produce a substantial reflection. At great depth, however, seismic velocities tend to be somewhat more homogenized than they are near the surface, because microcracks and pores are closed within a few kilometers of the surface, so that differences in seismic velocity from one rock type to another tend to be diminished. Coupled with the fact that boundaries are not often easily traced to known interfaces at the surface or in drill holes, the causes of deep reflections are not always clear. They may be from metasedimentary rocks, mylonite zones, layered intrusions, fluids, or combinations of these.

Where such boundaries can be related to known features, it has been found that any feature in the preceding list may explain the reflections, so that without some additional information, it is difficult to uniquely identify the lithology of specific reflectors. Nevertheless, whether or not geologic causes of specific reflectors can be determined, the patterns of reflectivity provide first-order geometric frameworks for interpretation.

15.6 THE IMPORTANCE OF REGIONAL PROFILES—LONGER, DEEPER, MORE DETAILED

In order to provide valuable information on the regional structure of the lithosphere, **seismic profiles** must be hundreds of kilometers long. Imagine trying to look into a dark room through a small hole with the illumination on your side of the hole. As the hole is increased in size, more of the light can penetrate through the aperture, and more of the reflected light from the objects inside the room is then visible from the vantage point outside the hole. Furthermore, as higher energy (brighter) light is used, the features within the darkened room become more visible. The situation is similar with seismic profiling: as longer profiles (apertures) are used, large-scale features are more likely to be seen, and as more energy (truck vibrator sources) is used, the input signal is larger, and more reflections are usually visible. By analogy, therefore, long seismic lines with many truck vibrators will be the most beneficial for mapping large structures of the lithosphere. Of course, these kinds of surveys require efforts that are correspondingly more expensive.

In addition to long profiles with large energy sources, it is desirable to obtain the most detailed geologic information possible. In order to accomplish this, the signal must include the widest possible range of frequencies. When much of the seismic profiling was initiated on regional scales across North America by the COCORP (COnsortium for COntinental Reflection Profiling) project from 1975–1980, technological limitations (long travel time for signals from the upper mantle) precluded acquisition of deep data with sufficiently high frequencies to provide much detail. Now there are close to 20,000 km that have been recorded in North America alone, and over the past 20 years technological developments have allowed acquisition of such extensive and detailed data, often with remarkable results. A data set from the LITHOPROBE program in Canada serves as an example to illustrate the approach to interpretation as well as some of the information that can be obtained.

15.7 AN EXAMPLE FROM NORTHWESTERN CANADA

A nearly 700-km long profile of the lithosphere in northwestern Canada recorded in 1996 and processed in 1996–1997 was acquired in an effort to map the deep structure of the western portion of the Canadian Shield, both where it is exposed on the east end of the profile and then where it projects beneath younger sedimentary rocks of the Western Canada Sedimentary Basin to the west (Figure 15.4). In this region, the Canadian Shield consists of the Archean Slave Province on the east, and younger, Proterozoic rocks on the west. The Proterozoic rocks are primarily associated with an orogen, the Wopmay Orogen, that has been interpreted from surface geologic measurements to represent remnants of tectonic accretion associated with subduction on the west margin of the Slave craton at about 1.85–2.1 Ga. On the west, the profile ended east of the Cordillera, although three more profiles that cross the Cordillera have since been recorded to provide nearly 3,000 km of data that extend from some of the oldest rocks in the world (Slave Province) to the modern active margin near Alaska.

The most dramatic features of this profile are the reflectivity throughout the crustal section, the regionally subhorizontal Moho, and the extensive, but comparatively sparse, upper mantle reflections (Figure 15.5). It is commonly observed that the crust is more reflective than the mantle and this profile is a nice illustration. This makes sense geologically because the crust is lithologically heterogeneous at many scales, including scales of tens to hundreds of meters, in which the seismic waves are most responsive. In contrast, the mantle tends to be more lithologically, and thus seismically, homogeneous.

The difficulties of interpreting the causes of crustal reflectivity are, however, amplified for mantle reflections because (1) mantle reflections cannot be linked directly to outcrop, and (2) the relatively homogeneous lithology of the mantle is not usually expected to have sufficient contrasts in properties to produce reflections. Nevertheless, reflections are present from within the mantle, and the large lateral extent of them implies that they are related to major regional features. Furthermore, even though the large-scale features are visible and mappable along this section, many smaller features, from the size of a sedimentary basin down to a few hundred meters, can also be delineated and are, indeed, most helpful in interpreting the large-scale structures.

It has been suggested that the regional patterns along this profile are related to Proterozoic subduction, with East-dipping mantle reflectors as images of a remnant subduction zone, and many of the crustal structures associated with this subduction and accretion process. The ability to image structures at various

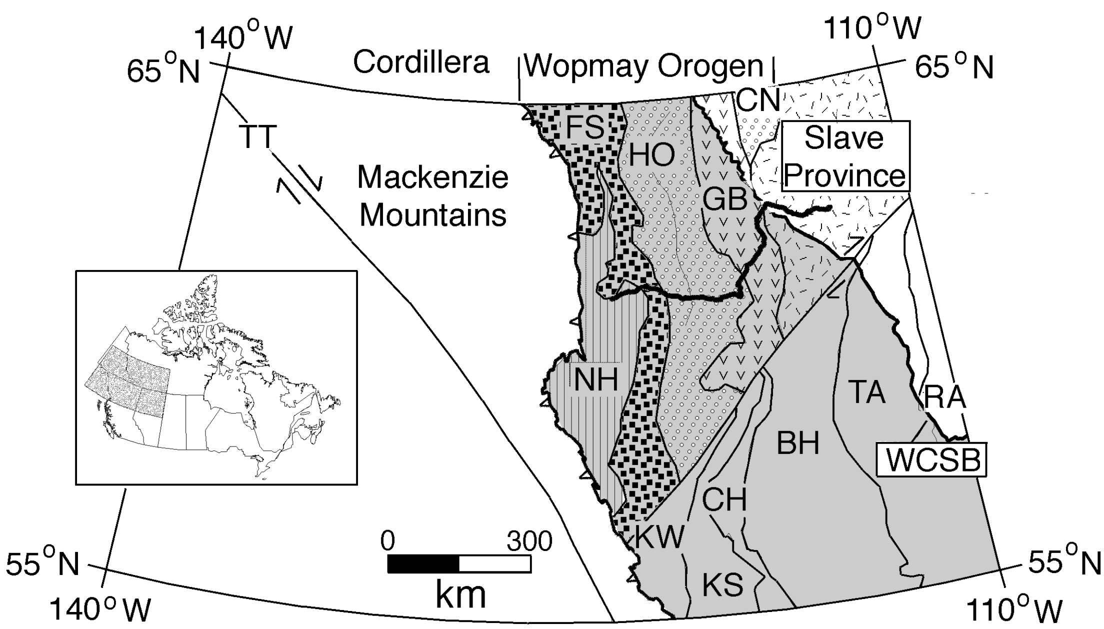

FIGURE 15.4 Map of northwestern Canada showing the division of major geologic domains and the location of a ~700-km long reflection profile (prominent dark line). Precambrian domains HO, GB, CN, RA, and TA are all defined on the basis of regional gravity and magnetic anomaly patterns that can be correlated to outcrops in the exposed Canadian Shield to the north and east. Domains NH, FS, KW, KS, CH, and BH are entirely covered by the sedimentary rocks of the Western Canada Sedimentary Basin (gray), the eastern edge of which is labeled WCSB.

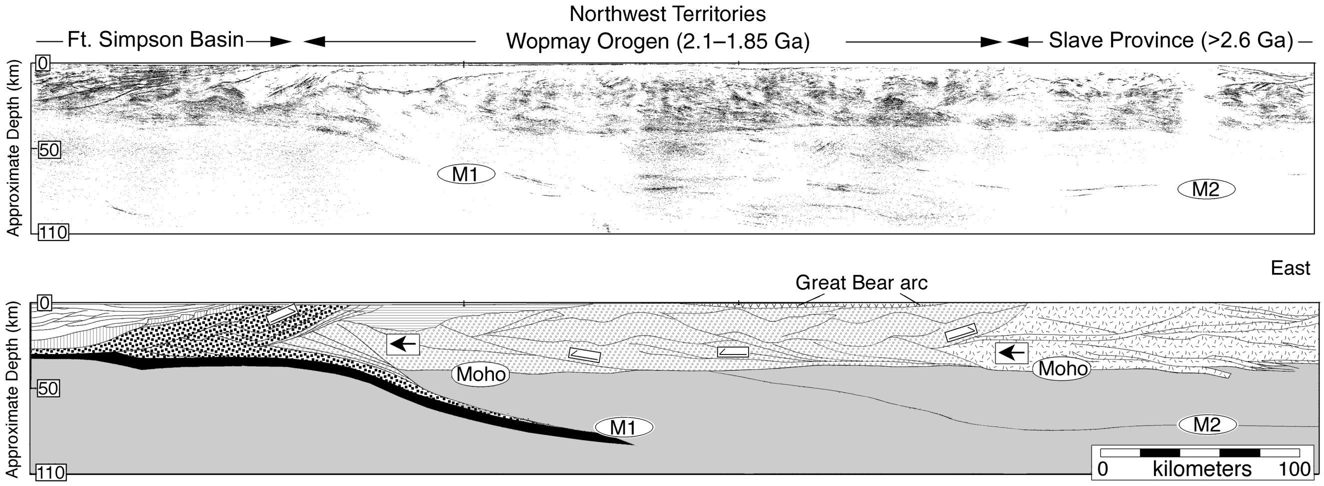

FIGURE 15.5 (upper) Regional seismic profile from ancient (>2.6 Ga) rocks of the Slave Province on the east, across the Proterozoic (2.1–1.85 Ga) Wopmay Orogen in the center, and then the younger Proterozoic (~ 1.74–0.55 Ga) Fort Simpson Basin on the west. The data are plotted to 32.0 s travel time, or about 120 km depth. Note the prominent crustal reflectivity, the crust–mantle transition, and sparse, but important reflections from within the upper mantle (M1 and M2). A general interpretation is shown (lower) to illustrate that the accretion of the Proterozoic rocks to the Slave Province probably resulted from subduction, the remnants of which are probably the dipping mantle reflections.

scales further allows the large and regionally significant boundaries to be related to more local features that, in turn, can be correlated with geologic observations (e.g., outcrop patterns, drill holes, and so on). For example, a key factor in the interpretation of the mantle reflections is that they can be followed to structures in the crust that can be approximately dated.

Consider the relationship between reflections M1 (Figure 15.5) and the crustal geometry previously discussed. On the west (left) side of the section, a series of layers thickens westward between the surface and 30 km depth (Figure 15.6a). These are almost certainly the expression of a westward thickening Proterozoic basin (the Fort Simpson Basin). They are overlain by shallow, more or less flat-lying, Paleozoic sedimentary rocks, and they are underlain by west-dipping surfaces that can be followed updip eastward to where they subcrop at the base of the Paleozoic. Drill holes have intersected the crystalline rocks that underlie the east flank of the basin and samples dated with radiometric techniques yield ages of about 1.845 Ga. Thus the basin layers overlying these crystalline rocks must be younger than about 1.845 Ga, but older than the Paleozoic.

Within the basin, even finer scale structures and stratigraphy may be discerned (Figure 15.6b). Near the surface, the unconformity between the base of the Paleozoic and the Proterozoic is evident as a truncation of dipping layers at ~0.2 s (about 500 m depth). Near here, drill holes penetrated from the Paleozoic sedimentary rocks into Proterozoic argillaceous rocks, thus establishing that the uppermost layers of the Proterozoic are indeed of sedimentary origin. Note also that the lower Paleozoic layers appear to be arched slightly into an anticline at the position of the truncation (Figure 15.6b). This anticline must have formed after the Paleozoic strata were deposited, and was probably associated with the uplift of the Cordillera, the eastern front of which is located about 50 km west of the profile. At greater reflection times, unconformities are visible within the Proterozoic layering, thus indicating that these deep layers are indeed also of sedimentary origin and filled a deep basin during this time (Figure 15.6a). Although the depth of the basin is not certain, the thickness of the layering indicates it may be as much as 20 km (Figure 15.6a). A prominent reflection crosses the stratified basin layers at a relatively high angle (Figure 15.6b). Although it is not known with certainty what this is, because it does not outcrop, its cross-cutting geometry is characteristic of dike intrusions, and there are such intrusions known within the Proterozoic sedimentary rocks of this region.

Thus, even though there are no direct observations (drill holes or outcrops) of the layering to 20 km depth, the large-scale geometry of a basin shape, the geometric relationships between the layers indicating unconformities and stratigraphic thickening (Figure 15.6a and b), and regional relationships that indicate Proterozoic sedimentary rocks are very thick to the west,

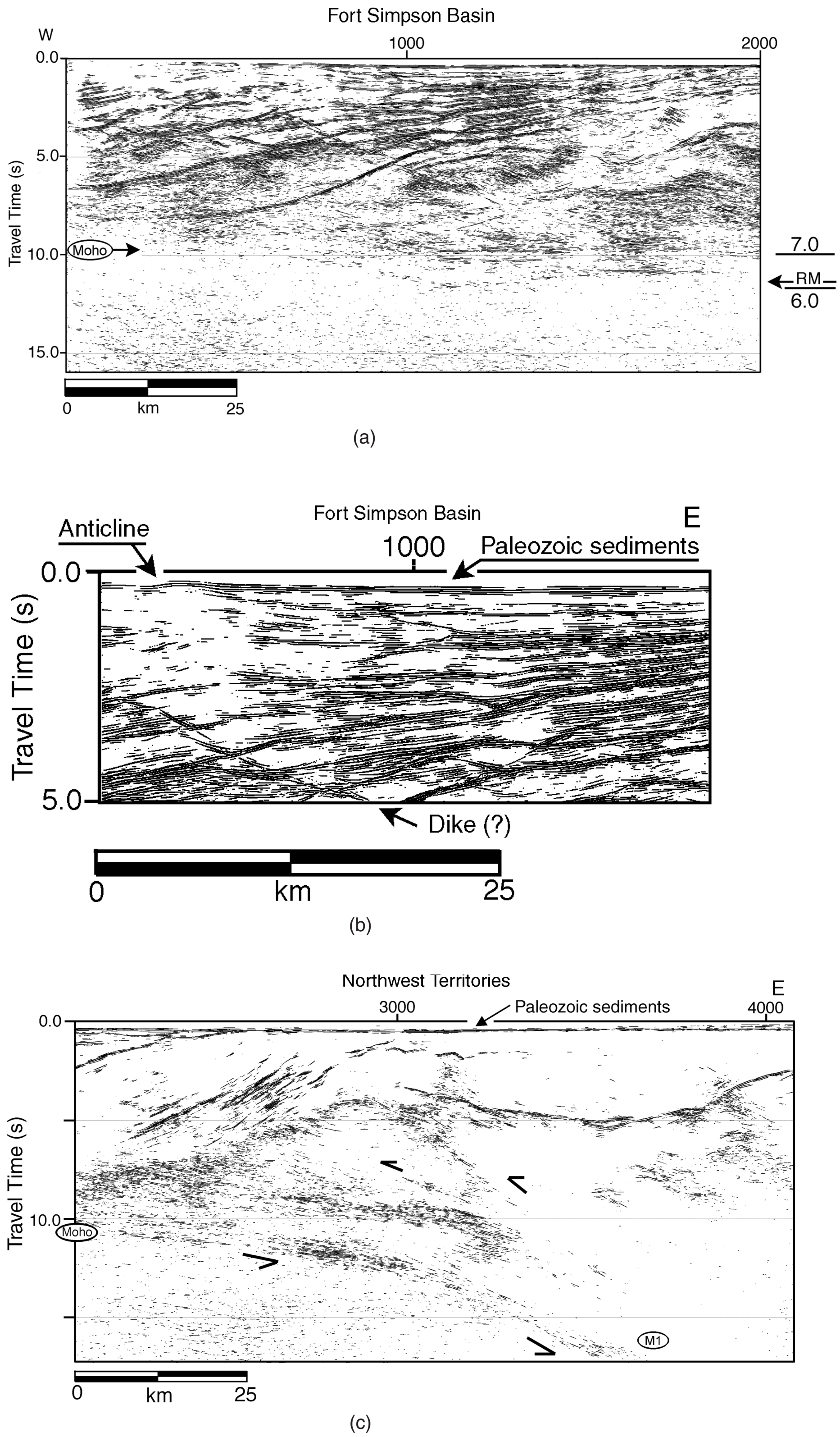

FIGURE 15.6 (a) The regional seismic profile across the Proterozoic basin illustrating the huge thickness of strata on the west and the associated shallowing of the Moho. (b) Enlargement of the regional profile in the upper part of the Proterozoic Fort Simpson Basin on the west. Note the sedimentary features such as the unconformity at the base of the Paleozoic sediments, unconformities in the eastward-thinning Proterozoic layers, and the prominent cross-cutting reflection that may be an igneous dike. (c) Enlargement of the regional profile across a feature that has been interpreted as the remnants of an accretionary complex. Note that the mantle reflections, M1, can be followed westward where they correlate with the Moho and that dipping layers above M1 tend to steepen eastward (upper arrows) as is common in accretionary wedges.

all lead to the conclusion that the western 100 km or so of this profile provides an image of a large, deep, and previously unknown sedimentary basin.

Reflections from layered sediments are well known in petroleum industry exploration. Thus, mapping such reflections from a large basin, even though it is Proterozoic, are not surprising. However, most of the deep continent includes crystalline metamorphic and igneous rock, and the common belief 20 to 25 years ago was that the velocity and density contrasts in these rocks were insufficient to produce reflections. If properly recorded and processed, however, data from the crystalline crust can have reflections that are just as prominent as those from a sedimentary basin.

To the east of the sedimentary basin discussed above, the crust beneath the thin Paleozoic cover consists of Proterozoic igneous and metamorphic rocks. This is known because the Paleozoic rocks thin to zero eastward (Figure 15.4), with crystalline rocks exposed on the surface east of there, and because drill holes intersect Precambrian crystalline rocks where the Paleozoic cover is present. Throughout this region, however, reflections are visible between the surface and about 35 km depth (Figure 15.5), which must all be within the crystalline basement.

The complex reflections from the crystalline crust can be interpreted by applying the same principles as with the Proterozoic basin: Drill holes provide direct evidence for the lithology and ages of rocks near the surface; the regional geology is incorporated to the extent possible, and detailed geometric relationships (i.e., truncations) are utilized to establish structural patterns and age relationships. Although there are too many details to address them completely here, three important characteristics stand out when the reflections are interpreted:

1. The reflectivity pattern delineates a series of complex structures associated with the accretion of middle Proterozoic rocks to an older Archean craton (Slave craton).
2. These structures are for the most part confined to the crust.
3. The base of the layering is remarkably abrupt at about 10–11 s (about 30–33 km) beneath both the Proterozoic and the Archean regions.

The first of these is significant because it provides key evidence on how subduction and accretion occurred during the Proterozoic (about 1.85–2.1 Ga in this region). From the geometric and geologic information, it appears that the products (what is visible today) of the tectonic process acting at that time are nearly identical to structures in modern subduction accretion zones. For example, east of the Fort Simpson Basin and above the mantle subduction reflections, the crustal geometry is nearly identical to that of an accretionary wedge and associated structures (Figure 15.6c).

The second and third characteristics emphasize the apparent structural (or at least reflection) differences between the crust and the mantle. The base of the crustal reflectivity (the "reflection Moho") is nearly horizontal along most of the profile east of the Fort Simpson Basin and is at a travel time that corresponds to the Moho identified from collocated regional seismic refraction data. This means that the crust–mantle transition here is either a zone of late intrusives (e.g., sills) that underlies the crust, or that it is a structural detachment zone. Examination of some detailed features with travel times near 10–11 s (about 30–33 km depth) provides information to distinguish between these possibilities (Figure 15.3b), as many of the layers in the lower crust beneath the Great Bear arc (Figure 15.5) of the Wopmay Orogen are listric (flatten) into the horizontal reflections near the Moho. Thus, the crust–mantle transition at this location is almost certainly a structural detachment rather than a layered intrusion zone. There are many profiles around the world that have images of lower crustal structures listric into the Moho, but the data must be of sufficiently high quality and have sufficiently fine detail for such subtle structures to be observed.

At some locations, in contrast to the listric structures just noted, the crust–mantle transition may have characteristics appropriate for an interpretation of sill-like intrusions; an example is visible beneath the Slave Province (Figure 15.7). Here, reflections project from the lower crust to below the Moho, and horizontal reflections at the Moho appear to cross cut them.

Other characteristics of the variable crust–mantle transition on this profile include the following: At two locations, reflections dip from the lower crust into the mantle; one of these corresponds to the interpreted Proterozoic subduction zone, and the other is located beneath the Archean craton. In some parts of the profile, the reflectivity of the crust–mantle transition is weak or nonexistent; whereas in others it is flat and prominent. At some locations the reflection Moho is flat, whereas beneath the Proterozoic basin it shallows by a few kilometers (Figure 15.6b). Thus, many of the variable characteristics of the crust–mantle transition that occur on deep reflection data around the world are visible along this single profile over relatively short lateral distances.

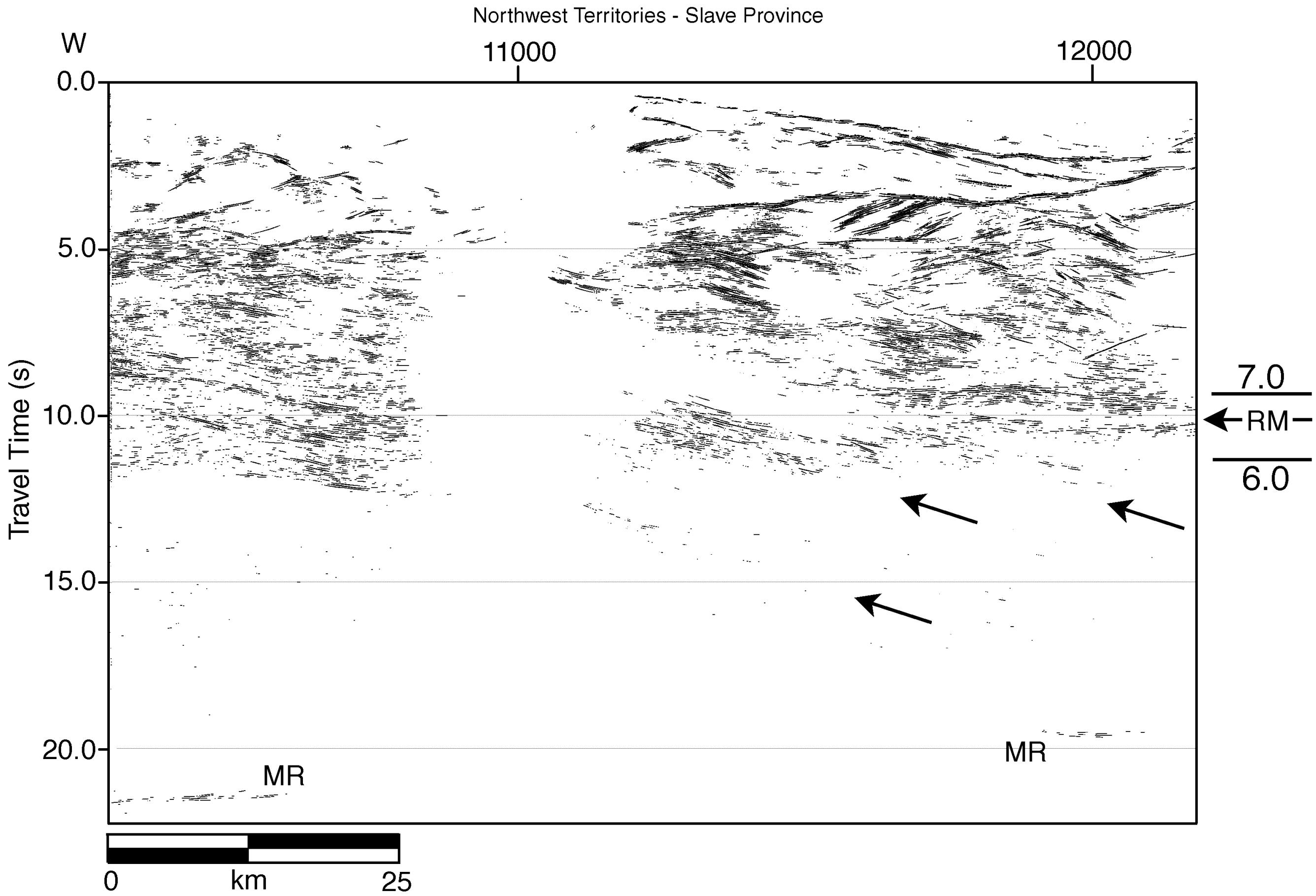

FIGURE 15.7 Enlargement of a segment of the regional profile from the Slave Province (Figure 15.5). Here, the Moho appears to have a series of dipping surfaces (arrows) that are cross cut by horizontal reflections (RM). One possible interpretation is that these horizontal reflections represent intrusives.

15.8 OTHER GEOPHYSICAL TECHNIQUES

Intracrustal structures such as the Proterozoic basin are sufficiently large to be compared with other regional data. The basin geometry on the west is observed on several other seismic profiles that are parallel to this one but that are located over a lateral distance of more than 500 km. They can be correlated from one to another with the application of other geophysical data; seismic profiles provide regional cross sections, but it is often difficult to project far away from the two-dimensional sections without additional information. In many areas, geophysical data such as gravity and magnetics, can provide such information.

The map in Figure 15.8a shows the isostatic gravity variations in northwestern Canada, and Figure 15.8b is an enlargement of the central portion of the map in the vicinity of the regional seismic profile. To produce this map, known characteristics of the Earth's shape and other effects have been estimated and removed from the measured values. The residual values, or anomalies, were contoured, and ideally represent variations due to rocks in the near subsurface. The contoured values were then plotted as a pseudotopographic image. As **gravity anomalies** result from variations in rock mass, in principle we should be able to determine the relative positions of different masses at depth. In practice, however, there is a fundamental problem underlying the interpretations of these results, as well as other geophysical anomalies such as magnetics. Because neither the subsurface structure nor the values of mass (or magnetism in the case of magnetic anomalies) of the rocks is known, an anomaly may be caused by small regions with large contrasts in properties, or large regions with small contrasts in properties. There are some limits (spatially), of course, because the anomalies are located according to map position (e.g., anomaly FS in Figure 15.8b must be due to something in the subsurface beneath it), but it is difficult to determine much more detail without some additional information from other techniques.

In the context of regional variations of continental structure, however, the patterns of large-scale anomalies can be extremely valuable in delineating patterns of continental structures. For example, in Figure 15.8a, the gravity patterns exhibit prominent, but relatively subdued, anomalies in the eastern part of the map (FS in Figure 15.8b); and more random and higher frequency patterns in the west, which are crossed by some major northwest oriented features (TT in Figure 15.8a). This change occurs where the sedimentary rocks of the Western Canada Sedimentary Basin give way to the complexly deformed rocks of the Mackenzie Mountains in the northern Cordillera. It is logical to interpret the change in gravity anomalies as being related to the large-scale geologic transition from the basin to the Cordillera.

On the other hand, the causes of the patterns beneath the Western Canada Basin are less clear. In this area, the sedimentary rocks are relatively flat and thin (Figure 15.6a), hence they should not exhibit major changes in gravity signature. The observed gravity variations

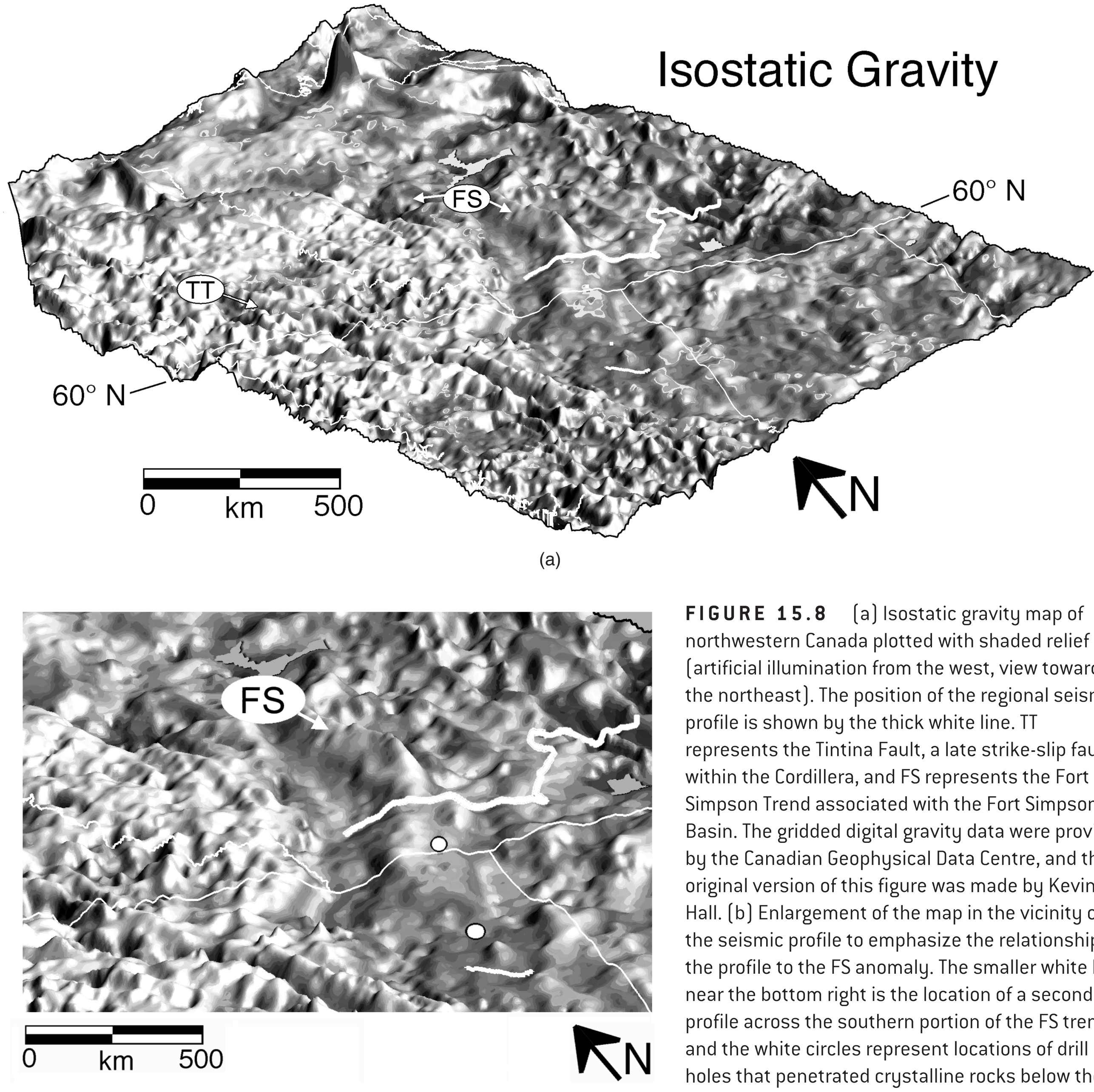

FIGURE 15.8 (a) Isostatic gravity map of northwestern Canada plotted with shaded relief (artificial illumination from the west, view toward the northeast). The position of the regional seismic profile is shown by the thick white line. TT represents the Tintina Fault, a late strike-slip fault within the Cordillera, and FS represents the Fort Simpson Trend associated with the Fort Simpson Basin. The gridded digital gravity data were provided by the Canadian Geophysical Data Centre, and the original version of this figure was made by Kevin Hall. (b) Enlargement of the map in the vicinity of the seismic profile to emphasize the relationship of the profile to the FS anomaly. The smaller white line near the bottom right is the location of a second profile across the southern portion of the FS trend, and the white circles represent locations of drill holes that penetrated crystalline rocks below the Western Canada Sedimentary Basin strata.

must therefore be associated with structure and lithology beneath the sedimentary basin, such as the large-scale structures observed on the reflection profile.

The interpretation of these observations is facilitated by the fact that the anomalies can be followed eastward into the Canadian Shield, where they are correlated with regional structures on the surface in the ancient (1.8–3.5 Ga) rocks. Accordingly, the patterns observed in the basin provide an image of structural patterns that project westward from the Canadian Shield, beneath the basin, to the eastern part of the Cordillera. Thus, while this approach does not necessarily provide us with much detail on the nature of the causes of individual anomalies, it does provide important information on the orientation and extent of subsurface structures in a region where there are no exposures.

All of the other seismic profiles that cross anomaly FS (Figure 15.8b) have essentially the same geometry; the Proterozoic basin previously described occurs everywhere to the west of FS. In all locations, the gravity signature indicates that the rocks at depth (below the Paleozoic sediments) are low density west of FS because the gravity anomalies are low, and this information is consistent with a deep, but old, basin within the crust. In applications for regional crustal and lithospheric imaging, therefore, one of the most valuable contributions of potential field maps is to project information over long distances away from the much higher resolution seismic cross sections.

15.9 CLOSING REMARKS

Geophysical imaging techniques have become standard tools for mapping the subsurface structure of the continental lithosphere. The most successful results derive from seismic reflection data that can be linked to known geologic features, either in outcrop or in drill holes. Improvements in field acquisition methods and signal processing have led to remarkable images of the crust and mantle lithosphere. Thus, as more, longer, and increasingly detailed profiles are acquired, the ability to compare, contrast, and link the results with other geologic and geophysical data will continue to foster new concepts on the origin and tectonic development of the continental lithosphere.

ADDITIONAL READING

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