CHAPTER FIVE

Rheology

5.1 INTRODUCTION

Earlier we defined strain as the shape change that a body undergoes in the presence of a stress field. But what do we really know about the corresponding stress? And is stress independent of strain? In this chapter we turn to the final and perhaps most challenging aspect of fundamental concepts: the relationship between stress and strain. Whereas it is evident that there is no strain without stress, the relationship between stress and strain is not easy to define on a physical basis. In other words, realizing that stress and strain in rocks are related is quite a different matter from physically determining their actual relationship(s). In materials science and geology we use the term **rheology** (from the Greek *rheos,* meaning "stream" or "flow") to describe the ability of stressed materials to deform or to flow, using fundamental parameters such as strain rate (strain per unit of time; Section 5.1.1), elasticity (Section 5.3.1), and viscosity (Section 5.3.2). These and other concepts will be discussed in this chapter, where we look especially at their significance for understanding rock deformation, rather than focusing on the associated mathematics. Recalling that stress and strain are second-order tensors, their proportionality is therefore a fourth-order tensor (that is, there are $3^4 = 81$ components). Up front we give a few brief, incomplete descriptions of the most important concepts that will appear throughout this chapter (Table 5.1) to help you to navigate through some of the initial material, until more complete definitions can be given.

Rheology is the study of flow of matter. Flow is an everyday phenomenon and in the previous chapter (Section 4.1) we used syrup on pancakes and human motion as day-to-day examples of deformation. Rocks don't seem to change much by comparison, but remember that geologic processes take place over hundreds of thousands to millions of years. For example, yearly horizontal displacement along the San Andreas Fault (a strike-slip fault zone in California) is on the order of a few centimeters, so considerable deformation has accumulated over the last 700,000 years. Likewise, horizontal displacements on the order of tens to hundreds of kilometers have occurred in the Paleozoic Appalachian fold-and-thrust belt of eastern North America over time period of a few million years (m.y.). Geologically speaking, time is available

TABLE 5.1 **BRIEF DESCRIPTIONS OF FUNDAMENTAL CONCEPTS AND TERMS RELATED TO RHEOLOGY**

Elasticity	Recoverable (non-permanent), instantaneous strain
Fracturing	Deformation mechanism by which a rock body or mineral loses coherency by simultaneously breaking many atomic bonds
Nonlinear viscosity	Permanent strain accumulation where the stress is exponentially related to the strain rate
Plasticity	Deformation mechanism that involves progressive breaking of atomic bonds without the material losing coherency
Strain rate	Rate of strain accumulation (typically, elongation, **e**, over time, t); shear strain rate, $\dot{\gamma}$ (gamma dot), is twice the longitudinal strain rate
Viscosity	Non-recoverable (permanent) strain that accumulates with time; the strain rate–stress relationship is linear

in large supply, and given sufficient amounts of it, rocks are able to flow, not unlike syrup. Glacier ice offers an example of flow in a solid material that shows relatively large displacements on human time scales (Figure 5.1). The flow of window glass, on the other hand, is an urban legend that you can refute with the information presented in this chapter.[1] When you look through the windows of an old house you may find that the glass distorts your view. The reason is, as the story goes, that the glass has sagged under its own weight with time (driven by gravity), giving rise to a wavy image. One also finds that the top part of the glass is often thinner than the bottom part. Using the viscous properties of glass (Table 5.5), however, you will see that this is likely due to old manufacturing processes rather than solid flow at surface temperatures. But before examining the mechanical behavior of materials, we need first to introduce the concept of strain rate.

5.1.1 Strain Rate

The time interval it takes to accumulate a certain amount of strain is described by the **strain rate,** symbol $\dot{e}$, which is defined as elongation (**e**) per time (t):[2]

$$\dot{e} = \mathbf{e}/t = \delta l/(l_o t) \qquad \text{Eq. 5.1}$$

You recall that elongation, length change divided by original length, $\delta l/l_o$ (Equation 4.2), is a dimensionless quantity; thus the dimension of strain rate is $[t]^{-1}$; the unit is second^{-1}. This may appear to be a strange unit at first glance, so let's use an example. If 30% finite longitudinal strain ($|\mathbf{e}| = 0.3$) is achieved in an experiment that lasts one hour (3600 s), the corresponding strain rate is $0.3/3600 = 8.3 \times 10^{-5}$/s. Now let's see what happens to the strain rate when we change the time duration of our experiment, while maintaining the same amount of finite strain.

Time interval for 30% strain	$\dot{e}$
1 day (86.4×10^3 s)	3.5×10^{-6}/s
1 year (3.15×10^7 s)	9.5×10^{-9}/s
1 m.y. (3.15×10^{13} s)	9.5×10^{-15}/s

Thus, the value of the strain rate changes as a function of the time period over which finite strain accumulates. Note that the percentage of strain did not differ for any of the time intervals. So what is the strain rate for a fault that moves 50 km in 1 m.y.? It is not possible to answer this question unless the displacement is expressed relative to another dimension of the body, that is, as a strain. We try again: What is the strain rate of an 800-km long fault moving 50 km in 1 m.y.? We get a strain rate of $(50/800)/(3.15 \times 10^{13}) = 2 \times 10^{-15}$/s. In many cases, commonly involving faults, geologists prefer to use the **shear strain rates** ($\dot{\gamma}$). The relationship between shear strain rate and (longitudinal) strain rate is

$$\dot{\gamma} = 2\dot{e} \qquad \text{Eq. 5.2}$$

A variety of approaches are used to determine characteristic strain rates for geologic processes. A widely used estimate is based on the Quaternary displacement along the San Andreas Fault of California, which gives a strain rate on the order of 10^{-14}/s. Other observations

---

[1] For an exposition of the legend, see the first edition of this book, p. 79.

[2] Here we consider only *longitudinal* strain rate.

FIGURE 5.1 Malaspina Glacier in Alaska showing moraines (dark bands) that are folded during differential flow of the ice.

(such as isostatic uplift, earthquakes,[3] and orogenic activity) support similar estimates and typical geologic strain rates therefore lie in the range of 10^{-12}/s to 10^{-15}/s. Now consider a small tectonic plate with a long dimension of 500 km at a divergent plate boundary. Using a geologic strain rate of 10^{-14}/s, we obtain the yearly spreading rate by multiplying this dimension of the plate by 3.15×10^{-7}/year, giving 16 cm/year, which is the order of magnitude of present-day plate velocities. On a more personal note, your 1.5-cm long fingernail grows 1 cm per year, meaning a growth rate of 0.67/year (or 2×10^{-8}/s). Your nail growth is therefore much, much faster than geologic rates, even though plates "grow" on the order of centimeters as well. We can offer many more geologic examples, but at this point we hope to leave you acquainted with the general concept of strain rate and typical values of 10^{-12}/s to 10^{-15}/s for geologic processes. Note that exceptions to this geologic range are rapid events like meteorite impacts and explosive volcanism, which are on the order of 10^{-2}/s to 10^{-4}/s.

5.2 GENERAL BEHAVIOR: THE CREEP CURVE

Compression tests on rock samples illustrate that the behavior of rocks to which a load is applied is not simple. Figure 5.2a shows what is called a **creep curve,** which plots strain as a function of time. In this experi-

---

[3] Remember that earthquakes, typically lasting only a few seconds, reflect discrete displacements (on the order of 10–100 cm); during periods in between there is no seismic activity and thus little or no slip.

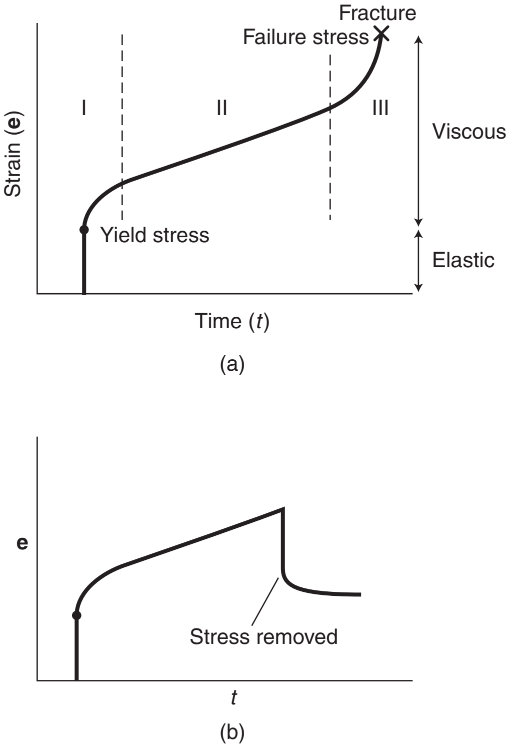


FIGURE 5.2 Generalized strain–time or creep curve, which shows primary (I), secondary (II), and tertiary (III) creep. Under continued stress the material will fail (a); if we remove the stress, the material relaxes, but permanent strain remains (b).

ment the differential stress is held constant. Three creep regimes are observed: (1) **primary** or **transient creep,** during which strain rate decreases with time following very rapid initial accumulation; (2) **secondary** or **steady-state creep,** during which strain accumulation is approximately linear with time; and (3) **tertiary** or **accelerated creep,** during which strain rate increases with time; eventually, continued loading will lead to failure. Restating these three regimes in terms of strain rate, we have regimes of (1) decreasing strain rate, (2) constant strain rate, and (3) increasing strain rate. The strain rate in each regime is the slope along the creep curve.

Rather than continuing our creep experiment until the material fractures, we decide to remove the stress sometime during the interval of steady-state creep. The corresponding creep curve for this second experiment is shown in Figure 5.2b. We see a rapid drop in strain when the stress is removed, after which the material relaxes a little more with time. Eventually there is no more change with time but, importantly, permanent strain remains. In order to examine this behavior of natural rocks we turn to simple analogies and rheologic models.

5.3 RHEOLOGIC RELATIONSHIPS

In describing the various rheologic relationships, we first divide the behavior of materials into two types, **elastic behavior** and **viscous behavior** (Figure 5.3). In some cases, the flow of natural rocks may be approximated by combinations of these **linear rheologies,** in which the ratio of stress over strain or stress over strain rate is a constant. The latter holds true for part of the mantle, but correspondence between stress and strain rate for many rocks is better represented by considering **nonlinear rheologies,** which we discuss after linear rheologies. For each rheologic model that is illustrated in Figure 5.3 we show a physical analog, a creep (strain–time) curve and a stress–strain or stress–strain rate relationship, which will assist you with the descriptions below. Such equations that describe the linear and nonlinear relationships between stress, strain, and strain rate are called **constitutive equations.**

5.3.1 Elastic Behavior

What is **elastic behavior** and is it relevant for deformed rocks? Let's first look at the relevance. In the field of seismology, the study of earthquakes, elastic properties are very important. As you know, seismic waves from an earthquake pass through the Earth to seismic monitoring stations around the world. As they travel, these seismic waves briefly deform the rocks, but after they have passed, the rocks return to their undeformed state. To imagine how rocks are able to do so we turn to a common analog: a rubber band. When you pull a rubber band, it extends; when you remove this stress, the band returns to its original shape. The greater the stress, the farther you extend the band. Beyond a certain point, called the failure stress, the rubber band breaks and brings a painful end to the experiment. This ability of rubber to extend lies in its atomic structure. The bond lengths between atoms and the angles between bonds in a crystal structure represent a state of lowest potential energy for a crystal. These bonds are able to elongate and change their relative angles to some extent, without introducing permanent changes in the crystal structure. Rubber bands extend particularly well because rubber can accommodate large changes in the angular relationships between bonds; however, this causes a considerable increase in the potential energy, which is recovered when we let go of the band, or when it snaps. So, once the stress is released, the atomic structure returns to its energetically most stable configuration, that is, the lowest potential energy. Like the elasticity of a rubber band, the ability of rocks to deform elastically also resides in nonpermanent distortions of

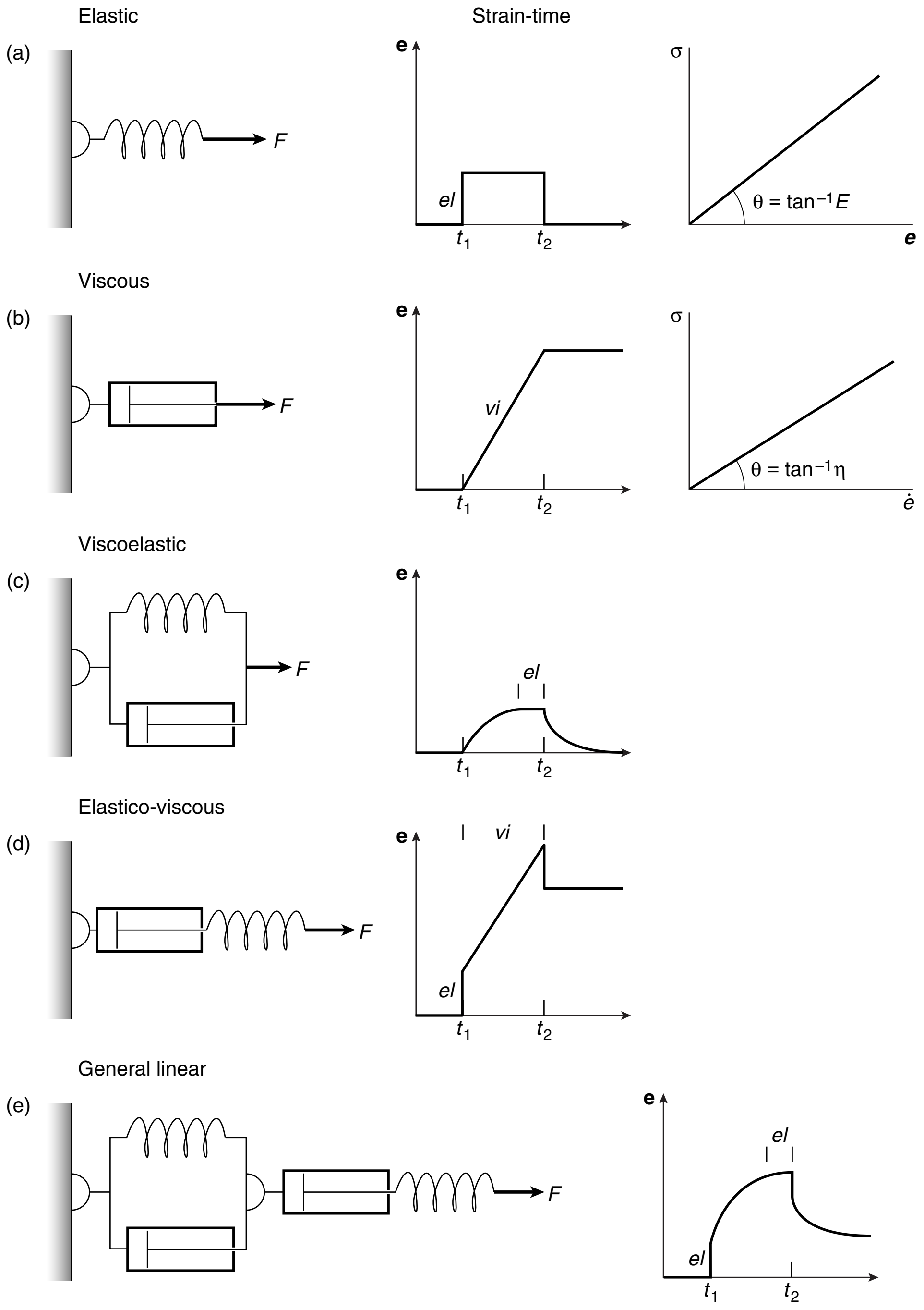


FIGURE 5.3 Models of linear rheologies. Physical models consisting of strings and dash pots, and associated strain–time, stress–strain, or stress–strain rate curves are given for (a) elastic, (b) viscous, (c) viscoelastic, (d) elastico-viscous, and (e) general linear behavior. A useful way to examine these models is to draw your own strain–time curves by considering the behavior of the spring and the dash pot individually, and their interaction. Symbols used: $\mathbf{e}$ = elongation, $\dot{e}$ = strain rate, σ = stress, *E* = elasticity, η = viscosity, *t* = time, *el* denotes elastic component, *vi* denotes viscous component.

the crystal lattice, but unlike rubber, the magnitude of this behavior is relatively small in rocks.

Expressing elastic behavior in terms of stress and strain, we get

$$\sigma = E \cdot \mathbf{e} \qquad \text{Eq. 5.3}$$

where E is a constant of proportionality called **Young's modulus** that describes the slope of the line in the σ–**e** diagram (tangent of angle θ; Figure 5.3a). The unit of this elastic constant is Pascal (Pa = kg/m · s^{-2}), which is the same as that of stress (recall that strain is a dimensionless quantity). Typical values of E for crustal rocks are on the order of -10^{11} Pa.[4] Linear Equation 5.3 is also known as **Hooke's Law,**[5] which describes elastic behavior. We use a spring as the physical model for this behavior (Figure 5.3a).

---

[4] We require a negative sign here to produce a negative elongation (shortening) from applying a compressive (positive) stress.

[5] After the English physicist Robert Hooke (1635–1703).

TABLE 5.2 SOME REPRESENTATIVE BULK MODULI (*K*) AND SHEAR MODULI (OR RIGIDITY, *G*) IN 10^{11} Pa AT ATMOSPHERIC PRESSURE AND ROOM TEMPERATURE

Crystal	*K*	*G*
Iron (Fe)	1.7	0.8
Copper (Cu)	1.33	0.5
Silicon (Si)	0.98	0.7
Halite (NaCl)	0.14	0.26
Calcite ($CaCO_3$)	0.69	0.37
Quartz (SiO_2)	0.3	0.47
Olivine (Mg_2SiO_4)	1.29	0.81
Ice (H_2O)	0.073	0.025

From Poirier (1985).

TABLE 5.3 SOME REPRESENTATIVE POISSON'S RATIOS (AT 200 MPa CONFINING PRESSURE)

Basalt	0.25
Gabbro	0.33
Gneiss	0.27
Granite	0.25
Limestone	0.32
Peridotite	0.27
Quartzite	0.10
Sandstone	0.26
Schist	0.31
Shale	0.26
Slate	0.30
Glass	0.24
Sponge	<<0.10

From Hatcher (1995) and other sources.

We can also write elastic behavior in terms of the shear stress, σ_s:

$$\sigma_s = G \cdot \gamma \qquad \text{Eq. 5.4}$$

where *G* is another constant of proportionality, called the **shear modulus** or the **rigidity,** and γ is the shear strain.

The corresponding constant of proportionality in volume change (dilation) is called the **bulk modulus,** *K:*

$$\sigma = K \cdot [(V - V_o)/V_o] \qquad \text{Eq. 5.5}$$

Perhaps more intuitive than the bulk modulus is its inverse, 1/*K,* which is the **compressibility** of a material. Representative values for bulk and shear moduli are listed in Table 5.2.

It is quite common to use an alternative to the bulk modulus that expresses the relationship between volume change and stress, called **Poisson's ratio,**[6] represented by the symbol ν. This elastic constant is defined as the ratio of the elongation perpendicular to the compressive stress and the elongation parallel to the compressive stress:

$$\nu = \mathbf{e}_{\text{perpendicular}}/\mathbf{e}_{\text{parallel}} \qquad \text{Eq. 5.6}$$

Poisson's ratio describes the ability of a material to shorten parallel to the compression direction without corresponding thickening in a perpendicular direction. Therefore the ratio ranges from 0 to 0.5, for fully compressible to fully incompressible materials, respectively. Incompressible materials maintain constant volume irrespective of the stress. A sponge has a very low Poisson's ratio, while a metal cylinder has a relatively high value. A low Poisson's ratio also implies that a lot of potential energy is stored when a material is compressed; indeed, if we remove the stress from a sponge it will jump right back to its original shape. Values for Poisson's ratio in natural rocks typically lie in the range 0.25–0.35 (Table 5.3).

A central characteristic of elastic behavior is its *reversibility:* once you remove the stress, the material returns to its original shape. Reversibility implies that the energy introduced remains available for returning the system to its original state. This energy, which is a form of potential energy, is called the **internal strain energy.** Because the material is undistorted after the stress is removed, we therefore say that strain is **recoverable.** Thus, elastic behavior is characterized by **recoverable strain.** A second characteristic of elastic behavior is the instantaneous response to stress: finite

---

[6] Named after the French mathematician Simeon-Denis Poisson (1781–1840).

TABLE 5.4	**ELASTIC CONSTANTS**
Bulk modulus (K)	Ratio of pressure and volume change
Compressibility ($1/K$)	The inverse of the bulk modulus
Elasticity (E)	Young's modulus
Poisson's ratio (ν)	A measure of compressibility of a material. It is defined as the ratio between **e** normal to compressive stress and **e** parallel to compressive stress.
Rigidity (G)	Shear modulus
Shear modulus (G)	Ratio of the shear stress and the shear strain
Young's modulus (E)	Ratio of compressive stress and longitudinal strain

strain is achieved immediately. Releasing the stress results in an instantaneous return to a state of no strain (Figure 5.3a). Both these elastic properties, recoverable and instantaneous strain, are visible in our rubber band or sponge experiments. However, elastic behavior is complicated by the granular structure of natural rocks, where grain boundaries give rise to perturbations from perfect elastic behavior in nongranular solids like glass. A summary of the elastic constants is given in Table 5.4.

Now we return to our original question about the importance of elastic behavior in rocks. With regard to finite strain accumulation, elastic behavior is relatively unimportant in naturally deformed rocks. Typically, elastic strains are less than a few percent of the total strain. So the answer to our question on the importance of elasticity depends on your point of view; a seismologist will say that elastic behavior is important for rocks, but a structural geologist will say that it is not very important.

5.3.2 Viscous Behavior

The flow of water in a river is an example of **viscous behavior** in which, with time, the water travels farther downstream. With this viscous behavior, strain accumulates as a function of time, that is, strain rate. We describe this relationship between stress and strain rate as

$$\sigma = \eta \cdot \dot{e} \qquad \text{Eq. 5.7}$$

where η is a constant of proportionality called the **viscosity** (tan θ, Figure 5.3b) and $\dot{e}$ is the strain rate. This ideal type of viscous behavior is commonly referred to as *Newtonian*[7] or **linear viscous behavior,** but do not confuse the use of "linear" in linear viscous behavior with that in linear stress–strain relationships in the previous section on elasticity. The term *linear* is used here to emphasize a distinction from nonlinear viscous (or non-Newtonian) behavior that we discuss later (Section 5.3.6).

To obtain the dimensional expression for viscosity, remember that strain rate has the dimension of $[t^{-1}]$ and stress has the dimension $[ml^{-1}t^{-2}]$. Therefore η has the dimension $[ml^{-1}t^{-1}]$. In other words, the SI unit of viscosity is the unit of stress multiplied by time, which is Pa · s (kg/m · s). In the literature we often find that the unit Poise[8] is used, where 1 Poise = 0.1 Pa · s.

The example of flowing water brings out a central characteristic of viscous behavior. Viscous flow is irreversible and produces **permanent** or **non-recoverable strain.** The physical model for this type of behavior is the dash pot (Figure 5.3b), which is a leaky piston that moves inside a fluid-filled cylinder.[9] The resistance encountered by the moving piston reflects the viscosity of the fluid. In the classroom you can model viscous behavior by using a syringe with one end open to the air. To give you a sense of the enormous range of viscosities in nature, the viscosities of some common materials are listed in Table 5.5.

How does the viscosity of water, which is on the order of 10^{-3} Pa · s (Table 5.5), compare with that of rocks? Calculations that treat the mantle as a viscous medium produce viscosities on the order of 10^{20}–10^{22} Pa · s. Obviously the mantle is much more viscous than water (>20 orders of magnitude!). You can demonstrate this graphically when calculating the slope of the lines for water and mantle material in the stress–strain rate diagram; they are 0.06° and nearly 90°, respectively. The much higher viscosity of rocks implies that motion is transferred over much larger distances. Stir water, syrup, and jelly in a jar to get a sense of this implication of viscosity. Obviously there is an enormous difference between materials that flow in our daily experience, such as water and syrup, and the "solids" that make up the Earth. Nevertheless, we can approximate the behavior of the Earth as a viscous

---

[7] Named after the British physicist Isaac Newton (1642–1727).

[8] Named after the French physician Jean-Louis Poiseuille (1799–1869).

[9] Just like us you have probably never heard of a dash pot until now, but an old V8-engine that uses equal amounts of oil and gas will also do.

TABLE 5.5 **REPRESENTATIVE VISCOSITIES (IN Pa · s)**

Air	10^{-5}
Water	10^{-3}
Olive oil	10^{-1}
Honey	4
Glycerin	83
Lava	$10–10^{4}$
Asphalt	10^{5}
Pitch	10^{9}
Ice	10^{12}
Glass	10^{14}
Rock salt	10^{17}
Sandstone slab	10^{18}
Asthenosphere (upper mantle)	$10^{20}–10^{21}$
Lower mantle	$10^{21}–10^{22}$

From several sources, including Turcotte and Schubert (1982).

medium over the large amount of time available to geologic processes (we will return to this with modified viscous behavior in Section 5.3.4). Considering an average mantle viscosity of 10^{21} Pa · s and a geologic strain rate of 10^{-14}/s, Equation 5.7 tells us that the differential (or flow) stresses at mantle conditions are on the order of tens of megapascals. Using a viscosity of 10^{14} Pa · s for glass, flow at atmospheric conditions produces a strain rate that is much too slow to produce the sagging effect that is ascribed to old windows (see Section 5.1).

5.3.3 Viscoelastic Behavior

Consider the situation in which the deformation process is reversible, but in which strain accumulation as well as strain recovery are delayed; this behavior is called **viscoelastic behavior.**[10] A simple analog is a water-soaked sponge that is loaded on the top. The load on the soaked sponge is distributed between the water (viscous behavior) and the sponge material (elastic behavior). The water will flow out of the sponge in response to the load and eventually the sponge will support the load elastically. For a physical model we place a spring (elastic behavior) and a dash pot (viscous behavior) in parallel (Figure 5.3c). When stress is applied, both the spring and the dash pot move simultaneously. However, the dash pot retards the extension of the spring. When the stress is released, the spring will try to return to its original configuration, but again this movement is delayed by the dash pot.

The constitutive equation for viscoelastic behavior reflects this addition of elastic and viscous components:

$$\sigma = E \cdot \mathbf{e} + \eta \cdot \dot{e} \qquad \text{Eq. 5.8}$$

5.3.4 Elastico-Viscous Behavior

Particularly instructive for understanding earth materials is **elastico-viscous**[11] **behavior,** where a material behaves elastically at the first application of stress, but then behaves in a viscous manner. When the stress is removed the elastic portion of the strain is recovered, but the viscous component remains. We can model this behavior by placing a spring and a dash pot in series (Figure 5.3d). The spring deforms instantaneously when a stress is applied, after which the stress is transmitted to the dash pot. The dash pot will move at a constant rate for as long as the stress remains. When the stress is removed, the spring returns to its original state, but the dash pot remains where it stopped earlier. The constitutive equation for this behavior, which is not derived here, is

$$\dot{e} = \dot{\sigma}/E + \sigma/\eta \qquad \text{Eq. 5.9}$$

where $\dot{\sigma}$ is the stress per time unit (i.e., stress rate).

When the spring is extended, it stores energy that slowly relaxes as the dash pot moves, until the spring has returned to its original state. The time taken for the stress to reach 1/*e* times its original value is known as the **Maxwell relaxation time,** where *e* is the base of natural logarithm (*e* = 2.718). Stress relaxation in this situation decays exponentially. The Maxwell relaxation time, t_M, is obtained by dividing the viscosity by the shear modulus (or rigidity):

$$t_M = \eta/G \qquad \text{Eq. 5.10}$$

---

[10] Also known as firmo-viscous or Kelvinian behavior, after the Irish-born physicist William Kelvin (1824–1907).

[11] Also called Maxwellian behavior after the Scottish physicist James C. Maxwell (1831–1879).

In essence the Maxwell relaxation time reflects the dominance of viscosity over elasticity. If t_M is high then elasticity is relatively unimportant, and vice versa. Because viscosity is temperature dependent, t_M can be expressed as a function of temperature. Figure 5.4 graphs this relationship between temperature and time for appropriate rock properties and shows that mantle rocks typically behave in a viscous manner (as a fluid). The diagram also suggests that crustal rocks normally fail by fracture (elastic field), but lower crustal rocks deform by creep as well. This discrepancy reflects the detailed properties of crustal materials and their nonlinear viscosities, as discussed later.

Maxwell proposed this model to describe materials that initially show elastic behavior, but given sufficient time display viscous behavior, which matches the behavior of Earth rather well. Recall that seismic waves are elastic phenomena (acting over short time intervals) and that the mantle is capable of flowing in a viscous manner over geologic time (acting over long time intervals). Taking a mantle viscosity of 10^{21} Pa · s and a rigidity of 10^{11} Pa, and assuming an olivine-dominated mantle (Table 5.2), we get a Maxwell relaxation time for the mantle of 10^{10} s, or on the order of 1000 years. This time agrees well with the uplift that we see following the retreat of continental glaciers after the last Ice Age, which resulted in continued uplift of regions like Scandinavia over thousands of years after the ice was removed.

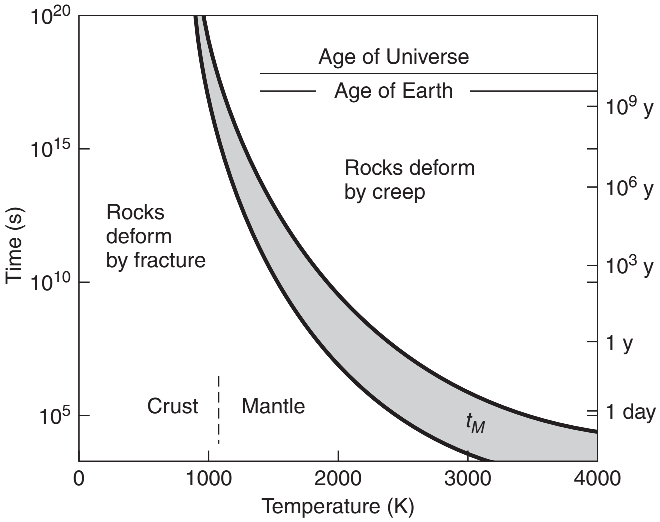


FIGURE 5.4 The Maxwell relaxation time, t_M, is plotted as a function of time and temperature. The curve is based on experimentally derived properties for rocks and their variabilities. The diagram illustrates that hot mantle rocks deform as a viscous medium (fluid), whereas cooler crustal rocks tend to deform by failure. This first-order relationship fits many observations reasonably well, but is incomplete for crustal deformation.

5.3.5 General Linear Behavior

So far we have examined two fundamental and two combined models and, with some further fine-tuning, we can arrive at a physical model that fairly closely approaches reality while still using linear rheologies. Such **general linear behavior** is modeled by placing the elastico-viscous and viscoelastic models in series (Figure 5.3e). Elastic strain accumulates at the first application of stress (the elastic segment of the elastico-viscous model). Subsequent behavior displays the interaction between the elastico-viscous and viscoelastic models. When the stress is removed, the elastic strain is first recovered, followed by the viscoelastic component. However, some amount of strain (permanent strain) will remain, even after long time intervals (the viscous component of the elastico-viscous model). The creep (**e**–*t*) curve for this general linear behavior is shown in Figure 5.3e and closely mimics the creep curve that is observed in experiments on natural rocks (compare with Figure 5.2b). We will not present the lengthy equation describing general linear behavior here, but you realize that it represents some combination of viscoelastic and elastico-viscous behavior.

5.3.6 Nonlinear Behavior

The fundamental characteristic that is common to all previous rheologic models is a linear relationship between strain rate and stress (Figure 5.5a): $\dot{e} \propto \sigma$. Experiments on geologic materials (like silicates) at elevated temperature show, however, that the relationship between strain rate and stress is often *nonlinear* (Figure 5.5b): $\dot{e} \propto \sigma^n$; where the exponent n is greater than 1. In other words, the proportionality of strain rate and stress is not constant. Rather, strain rate changes as a function of stress, and vice versa. This behavior is

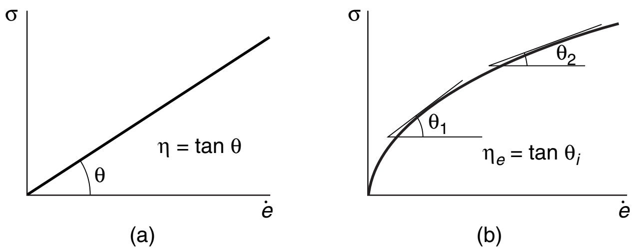


FIGURE 5.5 Linear (a) and nonlinear rheologies (b) in a stress–strain rate plot. The viscosity is defined by the slope of the linear viscous line in (a) and the effective viscosity by the slope of the tangent to the curve in (b).

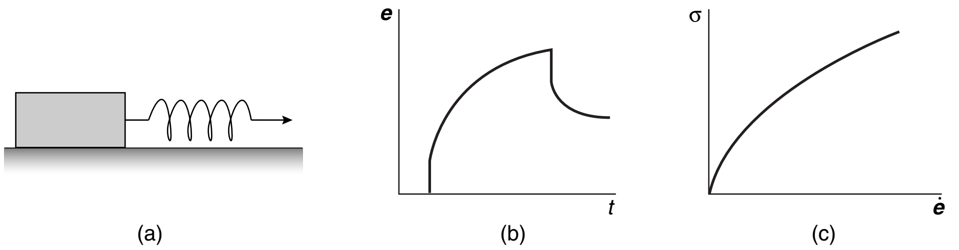


FIGURE 5.6 Nonlinear rheologies: elastic-plastic behavior. (a) A physical model consisting of a block and a spring; the associated strain–time curve (b) and stress–strain rate curve (c).

also displayed by wet paint, which therefore serves as a suitable analog. In order to understand the physical basis for nonlinear behavior in rocks we need to understand the processes that occur at the atomic scale during the deformation of minerals. This requires the introduction of many new concepts that we will not discuss here. We will return to this aspect in detail in Chapter 9, where we examine the role of crystal defects and their mobility.

A physical model representative of rocks showing nonlinear behavior is shown in Figure 5.6, and is known as **elastic-plastic behavior.** In this configuration a block and a spring are placed in series. The spring extends when a stress is applied, but only elastic (recoverable) strain accumulates until a critical stress is reached (the **yield stress**), above which the block moves and permanent strain occurs. The yield stress has to overcome the resistance of the block to moving (friction), but, once it moves, the stress remains constant while the strain accumulates. In fact, you experience elastic-plastic behavior when towing your car with a nylon rope that allows some stretch. Removing the elastic component (e.g., a sliding block with a nonelastic rope) gives **ideal plastic behavior,**[12] but this has less relevance to rocks.

A consequence of nonlinear rheologies is that we can no longer talk about (Newtonian) viscosity, because, as the slope of the stress–strain rate curve varies, the viscosity also varies. Nevertheless, as it is convenient for modeling purposes to use viscosity at individual points along the curve, we define the **effective viscosity** (η_e) as

$$\eta_e = \sigma/\dot{e} \qquad \text{Eq. 5.11}$$

This relationship is the same as that for viscous or Newtonian behavior (Equation 5.7), but in the case of effective viscosity you have to remember that η_e changes as the stress and/or the strain rate changes. In Figure 5.6 you see that the effective viscosity (the tangent of the slope) decreases with increasing stress and strain rate, which means that flow proceeds faster under these conditions. Thus, effective viscosity is not a material property like Newtonian viscosity, but a convenient description of behavior under prescribed conditions of stress or strain rate. For this reason, η_e is also called **stress-dependent** or **strain rate–dependent viscosity.**

The constitutive equation describing the relationship between strain rate and stress for nonlinear behavior is

$$\dot{e} = A\,\sigma^n \exp(-E^*/RT) \qquad \text{Eq. 5.12}$$

This relationship introduces several new parameters, some of which we used earlier without explanation (as in Figure 5.4). E^*, the activation energy, is an empirically derived value that is typically in the range of 100–500 kJ/mol, and n, the stress exponent, lies in the range $1 > n > 5$ for most natural rocks, with $n = 3$ being a representative value. The crystal processes that enable creep are temperature dependent and require a minimum energy before they are activated (see Chapter 9); these parameters are included in the exponential part of the function[13] as the activation energy (E^*) and temperature (T in degrees Kelvin); A is a constant, and R is the gas constant. Table 5.6 lists experimentally derived values for A, n, and E^* for many common rock types.

From Table 5.6 we can draw conclusions about the relative strength of the rock types, that is, their **flow stresses.** If we assume that the strain rate remains constant at 10^{-14}/s, we can solve the constitutive equation for various temperatures. Let us first rewrite Equation 5.12 as a function of stress:

$$\sigma = (\dot{e}/A)^{1/n} \exp(E^*/RT) \qquad \text{Eq. 5.13}$$

If we substitute the corresponding values for A, n, and E^* at constant T, we see that the differential stress value for rock salt is much less than that for any of the

---

[12] Or Saint-Venant behavior, after the French nineteenth-century physicist A. J. C. Barre de Saint-Venant.

[13] exp(a) means e^a, with e = 2.72.

TABLE 5.6 EXPERIMENTALLY DERIVED CREEP PARAMETERS FOR COMMON MINERALS AND ROCK TYPES

Rock type	A ($\mathrm{MPa}^{-n}\mathrm{s}^{-1}$)	n	E^* ($\mathrm{kJ \cdot mol^{-1}}$)
Albite rock	2.6×10^{-6}	3.9	234
Anorthosite	3.2×10^{-4}	3.2	238
Clinopyroxene	15.7	2.6	335
Diabase	2.0×10^{-4}	3.4	260
Granite	1.8×10^{-9}	3.2	123
Granite (wet)	2.0×10^{-4}	1.9	137
Granulite (felsic)	8.0×10^{-3}	3.1	243
Granulite (mafic)	1.4×10^{-4}	4.2	445
Marble ($<$ 20 MPa)	2.0×10^{-9}	4.2	427
Orthopyroxene	3.2×10^{-1}	2.4	293
Peridotite (dry)	2.5×10^{4}	3.5	532
Peridotite (wet)	2.0×10^{3}	4.0	471
Plagioclase (An75)	3.3×10^{-4}	3.2	238
Quartz	1.0×10^{-3}	2.0	167
Quartz diorite	1.3×10^{-3}	2.4	219
Quartzite	6.7×10^{-6}	2.4	156
Quartzite (wet)	3.2×10^{-4}	2.3	154
Rock salt	6.29	5.3	102

From Ranalli (1995) and other sources.

other rock types. This fits the observation that rock salt flows readily, as we learn from, for example, the formation of salt diapirs (Chapter 2). Limestones and marbles are also relatively weak and therefore regional deformation is often localized in these rocks. Quartz-bearing rocks, such as quartzites and granites, in turn are weaker than plagioclase-bearing rocks. Olivine-bearing rocks are among the strongest of rock types, meaning that they require large differential stresses to flow. But if this is true, how can the mantle flow at differential stresses of tens of megapascals? The answer is obtained by solving Equation 5.13 for various temperatures. In Figure 5.7 this is done using a "cold" geothermal gradient of 10 K/km[14] for some of the rock types in Table 5.6. The graph supports our first-order conclusion on the relative strength of various rock types, but it also shows that with increasing depth, the strength of all rock types decreases significantly. The latter is an important observation for understanding the nature of deformation processes in the deep Earth.

5.4 ADVENTURES WITH NATURAL ROCKS

Our discussion of rheology so far has been mostly abstract. We treated rocks as elastic springs and fluids, but we have not really looked at the behavior of natural rocks under different environmental conditions. The

---

[14] A representative "hot" geothermal gradient is 30 K/km.

results from decades of experiments on rocks will help us to get a better appreciation of the flow of rock. The reason for doing experiments on natural rocks is twofold: (1) we observe the actual behavior of rocks rather than that of syrup or elastic bands, and (2) we can vary several parameters in our experiments, such as pressure, temperature, and time, to examine their role in rock deformation. A vast amount of experimental data is available to us and many of the principles have, therefore, been known for several decades. Here we will limit our discussion by looking only at experiments that highlight particular parameters. By combining the various responses, we can begin to understand the rheology of natural rocks.

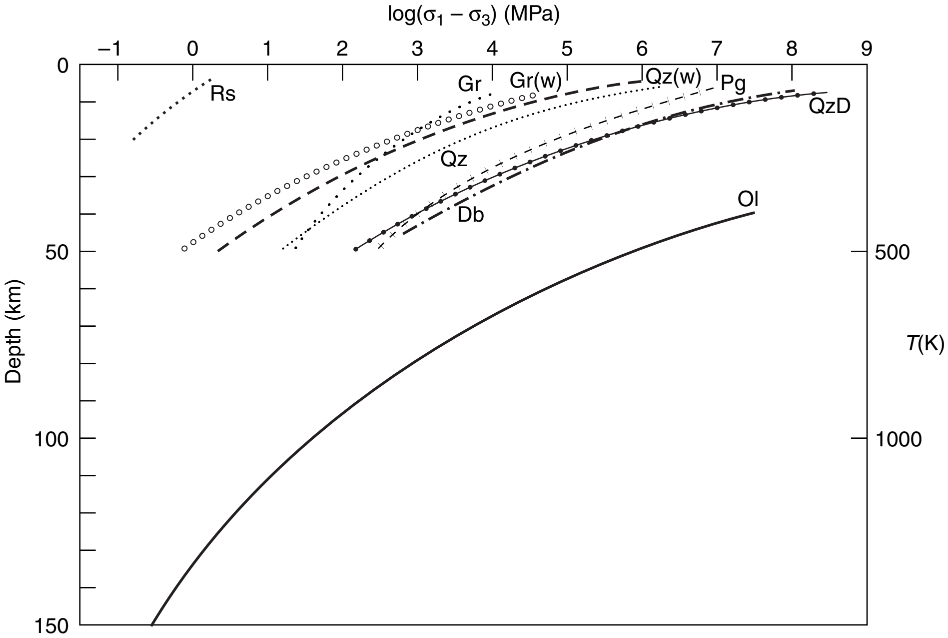


FIGURE 5.7 Variation of creep strength with depth for several rock types using a geothermal gradient of 10 K/km and a strain rate of 10^{-14}/s. Rs = rock salt, Gr = granite, Gr(w) = wet granite, Qz = quartzite, Qz(w) = wet quartzite, Pg = plagioclase-rich rock, QzD = quartz diorite, Db = diabase, Ol = Olivine-rich rock.

An alternative approach to examining the flow of rocks is to study material behavior in scaled experiments. Scaling brings fundamental quantities such as length [*l*], mass [*m*], and time [*t*] to the human scale. For example, we can use clay as a model material to study faulting, or wax to examine time-dependent behavior. Each analog that is used in scaled experiments has advantages and disadvantages, and the experimentalist has to make trade-offs between experimental conditions and geologic relevance. We will not use scaled experiments in the subsequent section of this chapter.

5.4.1 The Deformation Apparatus

A deformation experiment on a rock or a mineral can be carried out readily by placing a small sample in a vise, but when you try this experiment you have to be careful to avoid being bombarded with randomly flying chips as the material fails. If you ever cracked a hard nut in a nutcracker, you know what we mean. In rock deformation experiments we attempt to control the experiment a little better for the sake of the experimentalist, as well as to improve the analysis and interpretation of the results. A typical deformation apparatus is schematically shown in Figure 5.8. In this rig, a cylindrical rock specimen is placed in a pressure chamber, which is surrounded by a pressurized fluid that provides the **confining pressure,** P_c, on the specimen through an impermeable jacket. This experimental setup is known as a **triaxial testing apparatus,** named for the triaxial state of the applied stress, in which all three principal stresses are unequal to zero. For practical reasons, two of the principal stresses are equal. In addition to the fluid that provides the confining pressure, a second fluid may be present in the specimen to provide **pore-fluid pressure,** P_f. The difference between confining and pore pressure, $P_c - P_f$, is called the **effective pressure** (P_e). Adjusting the piston at the end of the test cylinder results in either a maximum or minimum stress along the cylinder axis, depending on the magnitudes of fluid pressure and axial stress. The remaining two principal stresses are equal to the effective pressure. By varying any or all of the axial stress, the confining pressure, or the pore-fluid pressure, we obtain a range of stress conditions to carry out our deformation experiments. In addition, we can heat the sample during the experiment to examine the effect of temperature. This configuration allows a limited range of finite strains, so a torsion rig with rotating plates is increasingly used for experiments at high shear strains.

A triaxial apparatus enables us to vary stress, strain, and strain rate in rock specimens under carefully controlled parameters of confining pressure, temperature, pore-fluid pressure, and time (that is, duration of the

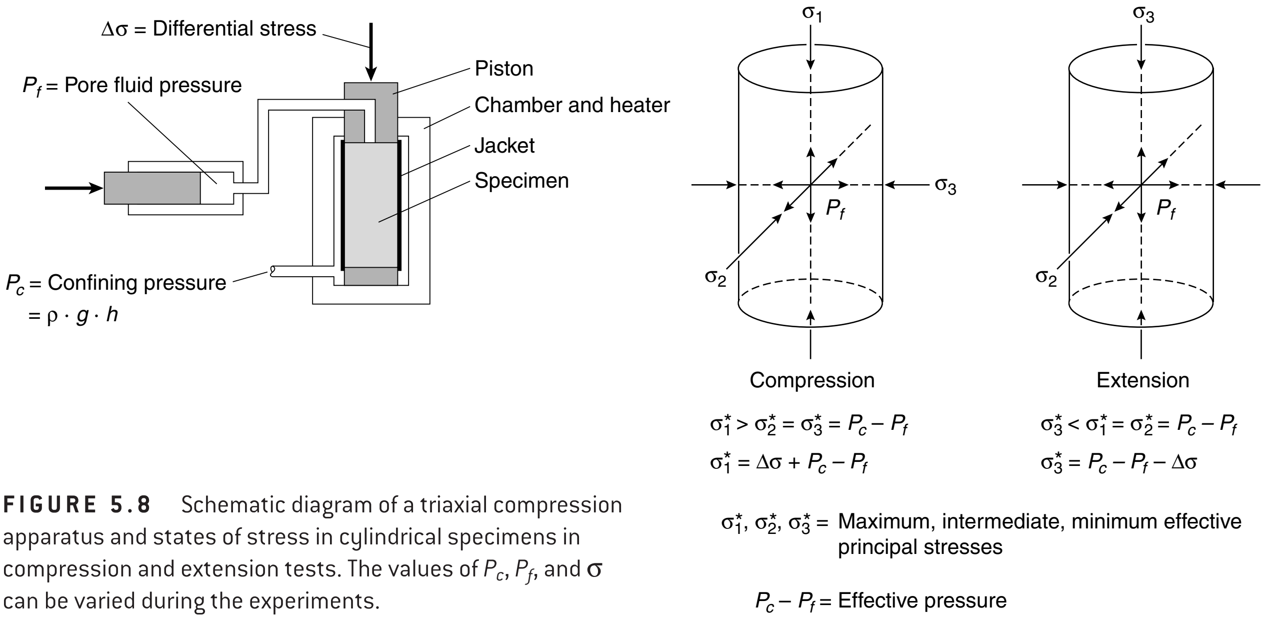


FIGURE 5.8 Schematic diagram of a triaxial compression apparatus and states of stress in cylindrical specimens in compression and extension tests. The values of P_c, P_f, and σ can be varied during the experiments.

experiment). What happens when we vary these parameters and what does this tell us about the behavior of natural rocks? Before we dive into these experiments it is useful to briefly review how these environmental properties relate to the Earth. Both confining pressure and temperature increase with depth in the Earth (Figure 5.9). The confining pressure is obtained from the simple relationship

$$P_c = \rho \cdot g \cdot h \qquad \text{(Eq. 4.15)}$$

where ρ is the density, g is gravity, and h is depth. This is the pressure from the weight of the overlying rock column, which we call the **lithostatic pressure.** The temperature structure of the Earth is slightly more complex than the constant gradient of 10 K/km used earlier in Figure 5.7. At first, temperature increases at an approximately constant rate (10°C/km–30°C/km),[15] but then the thermal gradient is considerably less (Figure 5.9). Additional complexity is introduced by the heat generated from compression at high pressures, which is reflected in the **adiabatic gradient** (dashed line in Figure 5.9). But if we limit our considerations to the crust and uppermost mantle, a linear **geothermal gradient** in the range of 10°C/km–30°C/km is a reasonable approximation. In Section 5.1 we learned that most geologic processes occur at strain rates on the order of 10^{-14}/s, with the exception of meteoric impacts, seismic events, and explosive volcanism. In contrast to geologic strain rates, experimental work is typically limited by the patience and the life expectancy of the experimentalist. Some of the slowest experiments are carried out at strain rates of 10^{-8}/s (i.e., 30% shortening in a year), which is still four to seven orders of magnitude greater than geologic rates. Having said all this, now let's look at the effects of varying environmental conditions, such as confining pressure, temperature, strain rate, and pore-fluid pressure, during deformation experiments.

5.4.2 Confining Pressure

You recall from Chapter 3 that confining pressure acts equally in all directions, so it imposes an isotropic stress on the specimen. When we change the **confining pressure** during our experiments we observe a very important characteristic: increasing confining pressure results in greater strain accumulation before failure (Figure 5.8). In other words, increasing confining pressure increases the viscous component and therefore the rock's ability to flow. What is the explanation for this? If you have already read Chapter 6, the Mohr circle for

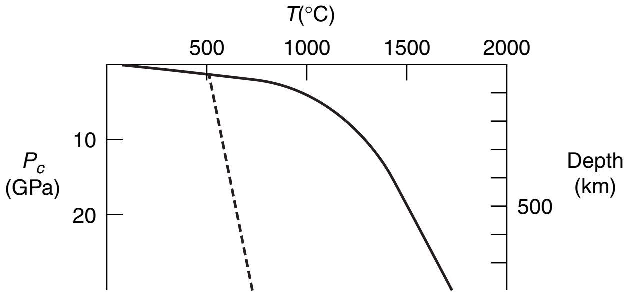
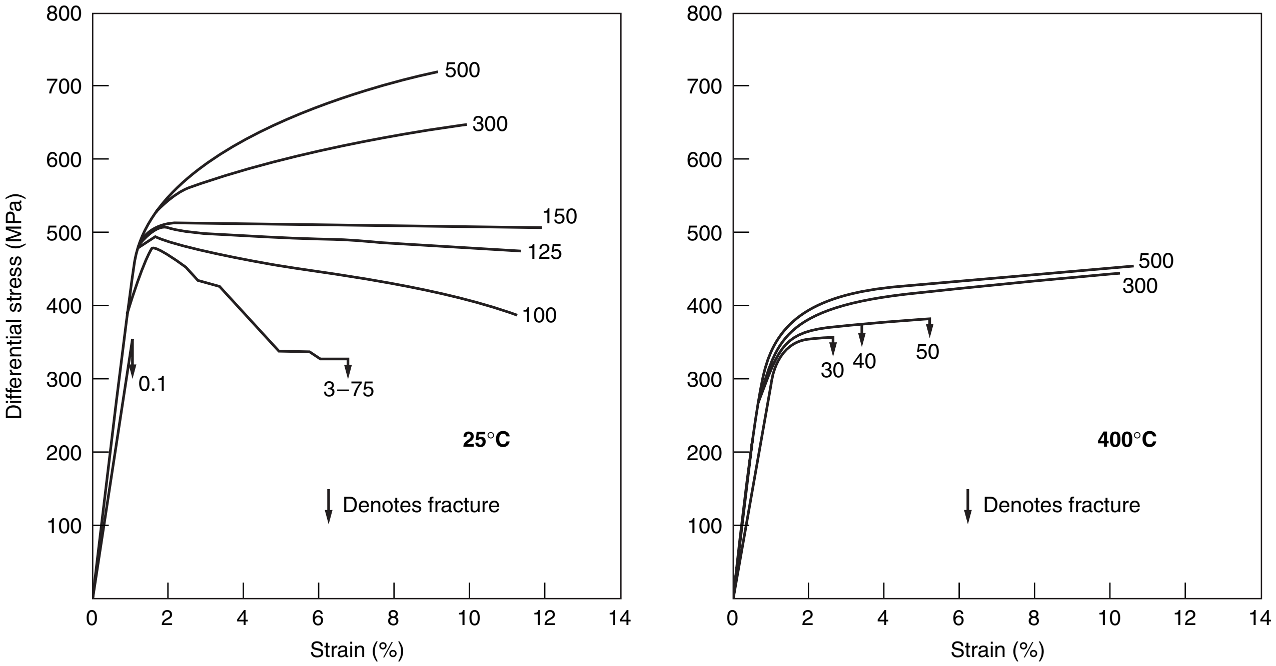


FIGURE 5.9 Change of temperature (T) and pressure (P_c) with depth. The dashed line is the adiabatic gradient, which is the increase of temperature with depth resulting from increasing pressure and the compressibility of silicates.

---

[15] Because we are dealing with rates, K and °C are interchangeable.



(a) (b)

FIGURE 5.10 Compression stress–strain curves of Solnhofen limestone at various confining pressures (indicated in MPa) at (a) 25°C and (b) 400°C.

stress and failure criteria give an explanation, but we will assume that this material is still to come. So, we'll take another approach. Moving your arm as part of a workout exercise is quite easy, but executing the same motion under water is a lot harder. Water "pushes" back more than air does, and in doing so it resists the motion of your arms. Similarly, higher confining pressures resist the opening of rock fractures, so any shape change that occurs is therefore viscous (ignoring the small elastic component).

The effect of confining pressure is particularly evident at elevated temperatures, where fracturing is increasingly suppressed (Figure 5.10b). When we compare common rock types, the role of confining pressure varies considerably (Figure 5.11); for example, the effect is much more pronounced in sandstone and shale than in quartzite and slate. Thus, it appears that larger strains can be achieved with increasing depth in the Earth, where we find higher lithostatic pressures.

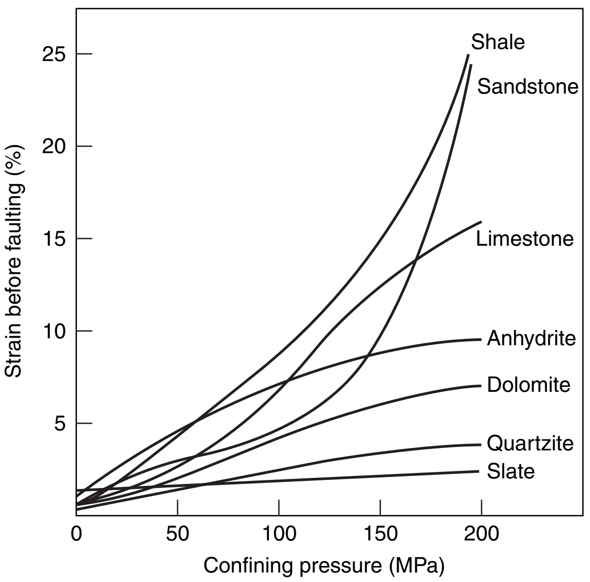
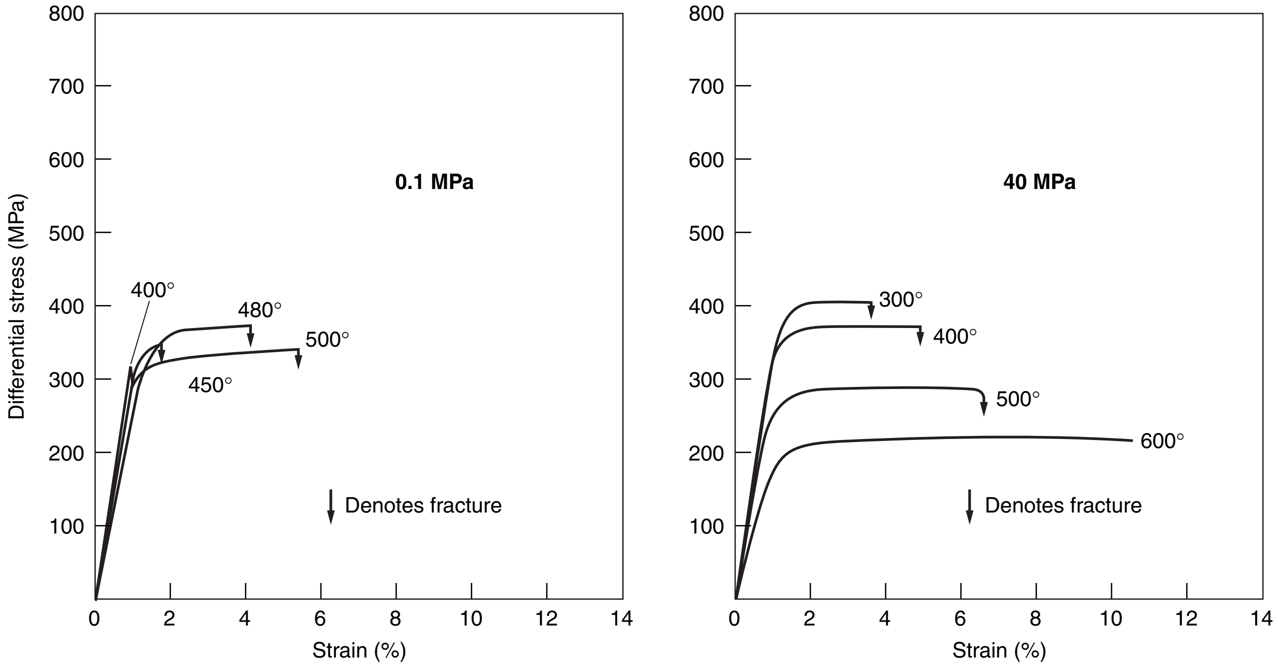


FIGURE 5.11 The effect of changing the confining pressure on various rock types. For these common rocks, the amount of strain before failure (ductility) differs significantly.

5.4.3 Temperature

A change in **temperature** conditions produces a marked change in behavior (Figure 5.12). Using the same limestone as in the confining pressure experiments (Figure 5.10), we find that the material fails rapidly at low temperatures. Moreover, under these conditions most of the strain prior to failure is recoverable (elastic). When we increase the temperature, the elastic portion of the strain decreases, while the ductility increases, which is most noticeable at elevated confining pressures (Figure 5.12b). You experience this temperature-dependence of flow also if you pour syrup on pancakes in a tent in the Arctic or in the Sahara: the ability of syrup to flow increases with temperature. Furthermore,



(a) (b)

FIGURE 5.12 Compression stress–strain curves of Solnhofen limestone at various temperatures (indicated in °C) at (a) 0.1 MPa confining pressure and at (b) 40 MPa confining pressure.

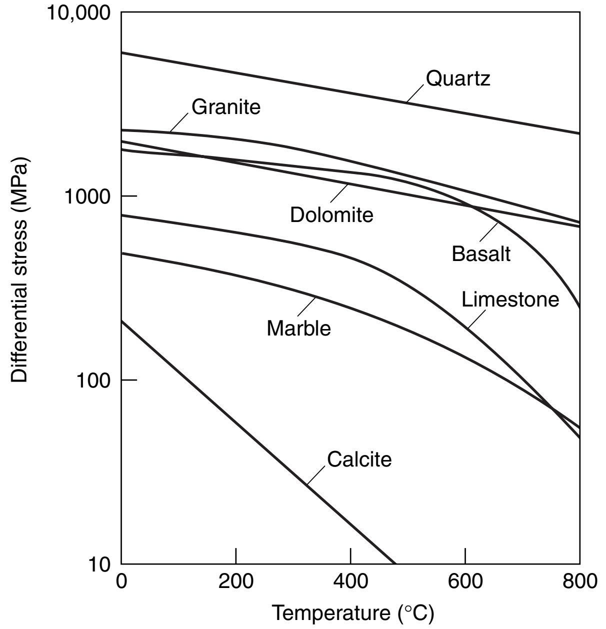


FIGURE 5.13 The effect of temperature changes on the compressive strength of some rocks and minerals.

the maximum stress that a rock can support until it flows (called the **yield strength** of a material) decreases with increasing temperature. The behavior of various rock types and minerals under conditions of increasing temperature is shown in Figure 5.13, from which we see that calcite-bearing rocks are much more affected than, say, quartz-bearing rocks. Collectively these experiments demonstrate that rocks have lower strength and are more ductile with increasing depth in the Earth, where we find higher temperatures.

5.4.4 Strain Rate

It is impossible to carry out rock deformation experiments at geologic rates (Section 5.4.1), so it is particularly important for the interpretation of experimental results to understand the role of **strain rate.** The effect is again best seen in experiments at elevated temperatures, such as those on marble (Figure 5.14). Decreasing the strain rate results in decreased rock strength and increased ductility. We again turn to an analogy for our understanding. If you slowly press on a small ball of Silly Putty®,[16] it spreads under the applied stress (ductile flow). If, on the other hand, you deform the same ball by a blow from a hammer, the material will shatter into many pieces (brittle failure). Although the environmental conditions are the same, the response is dramatically different because the strain rate differs.[17]

---

[16] A silicone-based material that offers hours of entertainment if you are a small child or a professional structural geologist.

[17] Assuming that the stress from a slow push and a rapid blow are equal.

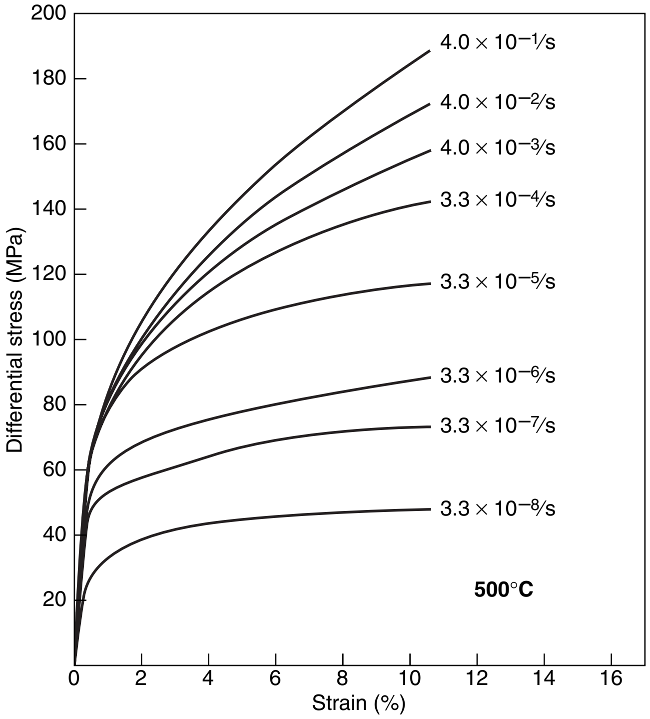


FIGURE 5.14 Stress versus strain curves for extension experiments in weakly foliated Yule marble for various constant strain rates at 500°C.

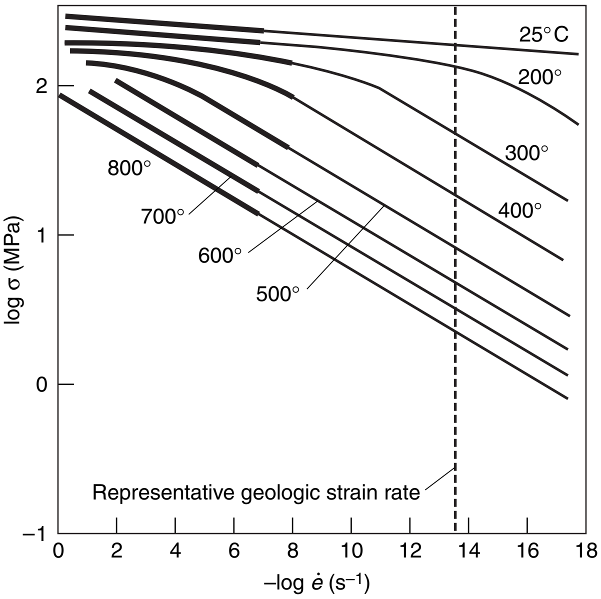


FIGURE 5.15 Log stress versus –log strain rate curves for various temperatures based on extension experiments in Yule marble. The heavy lines mark the range of experimental data; the thinner part of the curves is extrapolation to slower strain rates. A representative geologic strain rate is indicated by the dashed line.

Because rocks show similar effects from strain rate variation, the Silly Putty® experiment highlights a great uncertainty in experimental rock deformation. Extrapolating experimental results for strain rates over many orders of magnitude has significant consequences (Figure 5.15). Consider a strain rate of 10^{-14}/s and a temperature of 400°C, where ductile flow occurs at a differential stress of 20 MPa. At the same temperature, but at an experimental strain rate of 10^{-6}/s, the flow stresses are nearly an order of magnitude higher (160 MPa). Comparing this with the results of temperature experiments, you will notice that temperature change produces effects similar to strain rate variation in rock experiments (higher $t \propto$ lower $\dot{e}$); $\dot{e}$ has therefore been used as a substitute for geologic strain rates. In spite of the uncertainties, the volume of experimental work and our understanding of the mechanisms of ductile flow (Chapter 9) allow us to make reasonable extrapolations to rock deformation at geologic strain rates.

5.4.5 Pore-Fluid Pressure

Natural rocks commonly contain a fluid phase that may originate from the depositional history or may be secondary in origin (for example, fluids released from prograde metamorphic reactions). In particular, low-grade sedimentary rocks such as sandstones and shales, contain a significant fluid component that will affect their behavior under stress. To examine this parameter, the deformation rig shown in Figure 5.8 contains an impermeable jacket around the sample. Experiments show that increasing the **pore-fluid pressure** produces a drop in the sample's strength and reduces the ductility (Figure 5.16a). In other words, rocks are weaker when the pore-fluid pressure is high. We return to this in Chapter 6, but let us briefly explore this effect here. Pore-fluid pressure acts equally in all directions and thus counteracts the confining pressure, resulting in an effective pressure ($P_e = P_c - P_f$) that is less than the confining pressure. Thus, we can hypothesize that increasing the pore-fluid pressure has the same effect as decreasing the confining pressure of the experiment. We put this to the test by comparing the result of two experiments on limestone in Figure 5.16a and 5.16b that vary pore-fluid pressure and confining pressure, respectively. Clearly, there is remarkable agreement between the two experiments, supporting our hypothesis.

The role of fluid content is a little more complex than is immediately apparent from these experiments,

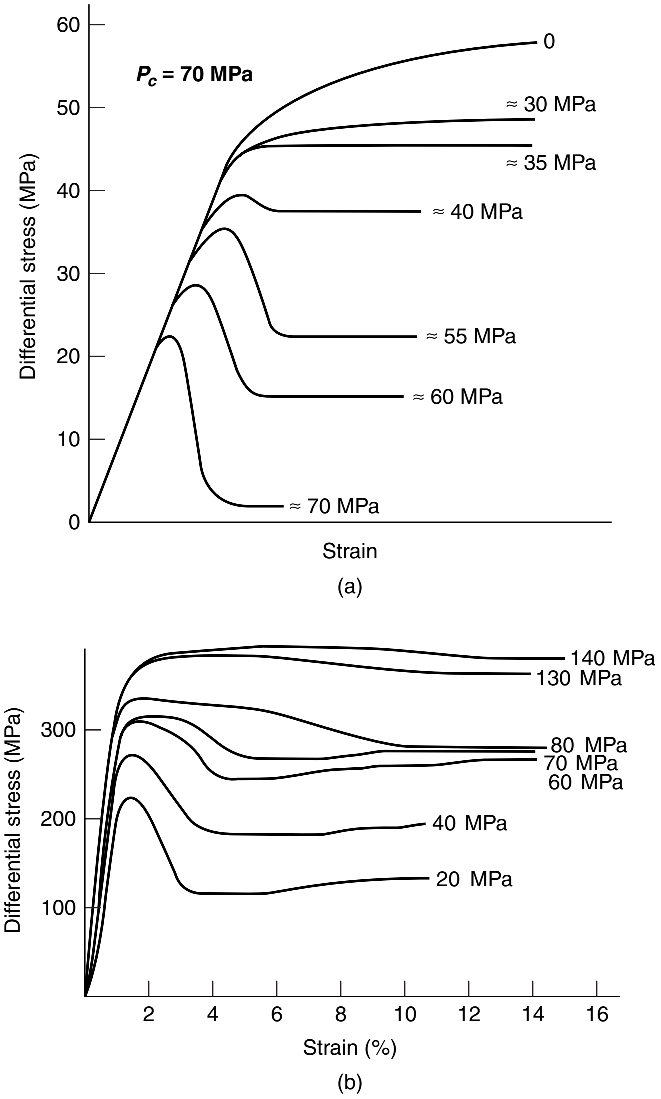


FIGURE 5.16 Comparing the effect on the behavior of limestone of (a) varying pore-fluid pressure and (b) varying confining pressure.

because of fluid-chemical effects. While ductility decreases with increasing pore-fluid pressure, the corresponding decreased strength of the material will actually promote flow. The same material with low fluid content ("dry" conditions) would resist deformation, but at high fluid content ("wet" conditions) flow occurs readily. This is nicely illustrated by looking at the deformation of quartz with varying water content (Figure 5.17). The behavior of quartz is similar to that in the previous rock experiment: the strength of "wet" quartz is only about one tenth that of "dry" quartz at the same temperature. The reason for this weakening lies in the substitution of OH groups for O in the silicate crystal lattice, which strains and weakens the

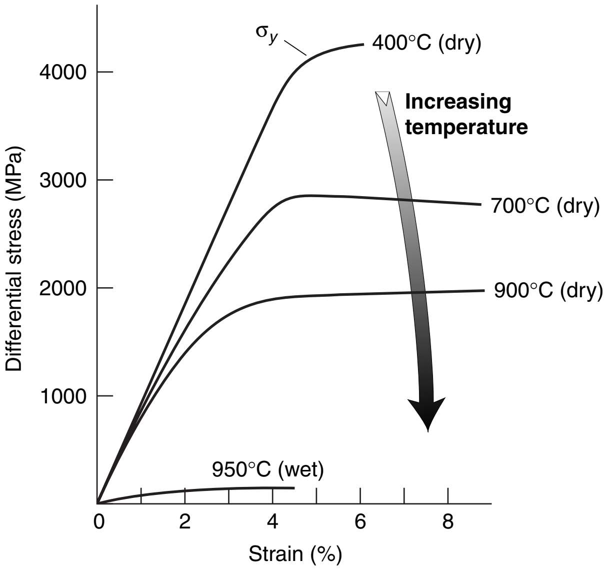


FIGURE 5.17 The effect of water content on the behavior of natural quartz. Dry and wet refer to low and high water content, respectively. The curves also show the effect of temperature on single crystal deformation, which is similar to that for rocks (Figure 5.12).

Si-O atomic bonds (see Chapter 6).[18] In practice, fluid content explains why many minerals and rocks deform relatively easily even under moderate stress conditions.

5.4.6 Work Hardening–Work Softening

Laboratory experiments on rocks bring an interesting property to light that we first noticed in the general creep curve of Figure 5.2a, namely, that the relationship between strain and time varies in a single experiment. The strain rate may decrease, increase, or remain constant under constant stress. When we carry out experiments at a constant strain rate we often find that the stresses necessary to continue the deformation may increase or decrease, phenomena that engineers call **work hardening** (greater stress needed) and **work softening** (lower stress needed), respectively. In a way you can think of this as the rock becoming stronger or weaker with increasing strain; therefore, we also call this effect **strain hardening** and **strain softening,** respectively. A practical application of work hardening is the repeated rolling of metal that gives it greater strength, especially when it is heated. While you may not realize this when seeing the repair bill after an

---

[18] This is called **hydrolytic weakening** (Chapter 6).

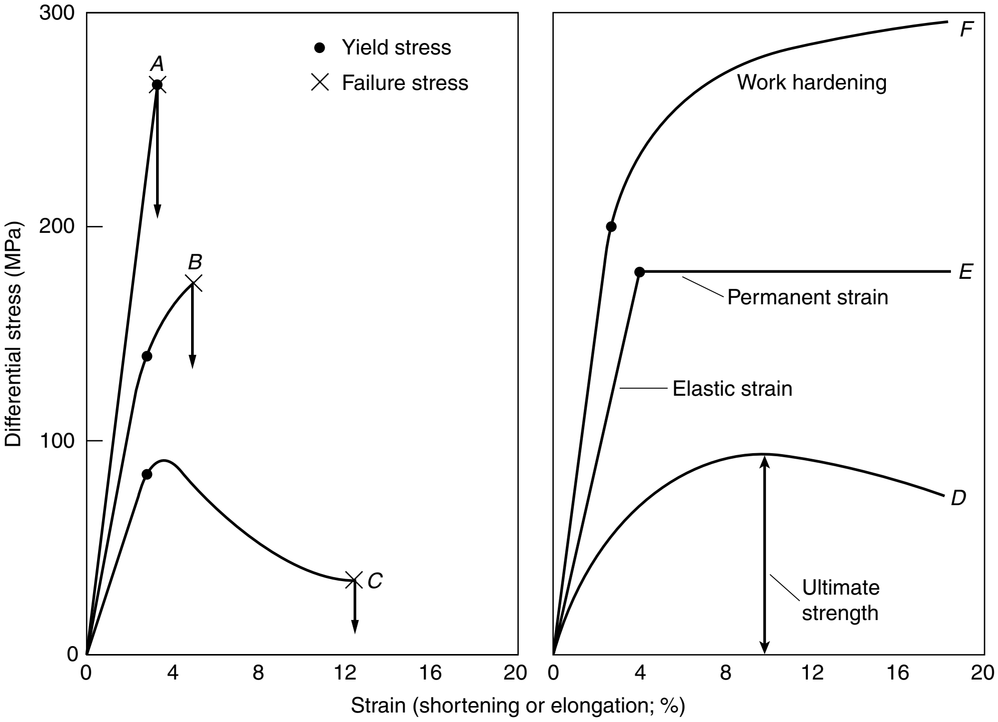


FIGURE 5.18 Representative stress–strain curves of brittle (*A* and *B*), brittle-ductile (*C*), and ductile behavior (*D–F*). *A* shows elastic behavior followed immediately by failure, which represents brittle behavior. In *B,* a small viscous component (permanent strain) is present before brittle failure. In *C,* a considerable amount of permanent strain accumulates before the material fails, which represents transitional behavior between brittle and ductile. *D* displays no elastic component and work softening. *E* represents ideal elastic-plastic behavior, in which permanent strain accumulates at constant stress above the yield stress. *F* shows the typical behavior seen in many of the experiments, which displays a component of elastic strain followed by permanent strain that requires increasingly higher stresses to accumulate (work hardening). The yield stress marks the stress at the change from elastic (recoverable or nonpermanent strain) to viscous (nonrecoverable or permanent strain) behavior; failure stress is the stress at fracturing.

unfortunate encounter with a moving tree, car makers use this to strengthen some metal parts. This effect was already known in ancient times, when Japanese samurai sword makers were producing some of the hardest blades available using repeated heating and hammering of the metal. Work softening is the opposite effect, in which the stress required to continue the experiment is less; in constant stress experiments this results in a strain rate increase. Both processes may occur in the same materials as shown in, for example, Figure 5.10a. At low confining pressure a stress drop is observed after about 2% strain, representing work softening. At high confining pressures, the limestone displays work hardening, because increasingly higher stresses are necessary to continue the experiment. Schematically this is summarized in Figure 5.18, where curve *D* displays work softening and curve *F* displays work hardening.

Work hardening and work softening are understood from the atomic-scale processes that enable rocks to flow. However, unless you already have some background in crystal plasticity a quick explanation here would be insufficient. We therefore defer explanation of the physical basis of these processes until the time we discuss crystal defects and their movement (Chapter 8).

5.4.7 Significance of Experiments to Natural Conditions

We close this section on experiments with a table that summarizes the results of varying the confining pressure, temperature, fluid pressure, and strain rate (Table 5.7) in experiments on rocks and by examining the significance for geologic conditions.

From Table 5.7 we see that increasing the confining pressure (P_c) and the fluid pressure (P_f) have

TABLE 5.7 EFFECT OF ENVIRONMENTAL PARAMETERS ON RHEOLOGIC BEHAVIOR

	Effect	Explanation
High P_c	Suppresses fracturing; increases ductility; increases strength; increases work hardening	Prohibits fracturing and frictional sliding; higher stress necessary for fracturing exceeds that for ductile flow
High T	Decreases elastic component; suppresses fracturing; increases ductility; reduces strength; decreases work hardening	Promotes crystal plastic processes
Low $\dot{e}$	Decreases elastic component; increases ductility; reduces strength; decreases work hardening	Promotes crystal plastic processes
High P_f	Decreases elastic component; promotes fracturing; reduces ductility; reduces strength *or* promotes flow	Decreases P_c ($P_e = P_c - P_f$) and weakens Si-O atomic bonds

opposing effects, while increasing temperature (T) and lowering strain rate ($\dot{e}$) have the same effect. Confining pressure and temperature, which both increase with depth in the Earth, increasingly resists failure, while promoting larger strain accumulation; that is, they increase the ability of rocks to flow. High pore-fluid content is more complicated as it favors fracturing if P_f is high or promotes flow in the case of intracrystalline fluids. From these observations we would predict that brittle behavior (fracturing) is largely restricted to the upper crust, while ductile behavior (flow) dominates at greater depth. A natural test supporting this hypothesis is the realization that faulting and earthquakes generally occur at shallow crustal levels (<15 km depth),[19] while large-scale ductile flow dominates the deeper crust and mantle (e.g., mantle convection).

5.5 CONFUSED BY THE TERMINOLOGY?

By now perhaps a baffling array of terms and concepts have passed before your eyes. If you think that you are the only one who is confused, you should look at the scientific literature on this topic. Let us, therefore, try to bring some additional order to the terminology. Table 5.8 lists brief descriptions of terms that are commonly used in the context of rheology, and contrasts the mechanical behavior of rock deformation with operative deformation mechanisms. A schematic diagram of some representative stress–strain curves summarizes the most important elements (Figure 5.18).

Perhaps the two most commonly used terms in the context of rheology are brittle and ductile. In fact, we use them to subdivide rock structures in Parts B and C of this text, so it is important to understand their meaning. **Brittle behavior** describes deformation that is localized on the mesoscopic scale and involves the formation of fractures. For example, a fracture in a tea cup is brittle behavior. In natural rocks, brittle fracturing occurs at finite strains of 5% or less. In contrast, **ductile behavior** describes the ability of rocks to accumulate significant permanent strain with deformation that is distributed on the mesoscopic scale. The shape change that we achieved by pressing our clay cube is ductile behavior. In nature, a faulted rock and a folded rock are examples of brittle and ductile behavior, respectively. Importantly, these two modes of behavior do not define the mechanism by which deformation occurs. This distinction between behavior and mechanism is important, and can be explained with a simple example.

Consider a cube that is filled with small undeformable spheres (e.g., marbles) and a second cube that is filled with spheres consisting of clay (Figure 5.19). If we deform these cubes into rectangular blocks, the mechanism by which this shape change is achieved is

---

[19] Excluding deep earthquakes in subduction zones, which represent special conditions.

TABLE 5.8 TERMINOLOGY RELATED TO RHEOLOGY, WITH EMPHASIS ON BEHAVIOR AND MECHANISMS

Brittle-ductile transition	Depth in the Earth below which brittle behavior is replaced by ductile processes (see under "behavior" below)
Brittle-plastic transition	Depth in the Earth where the dominant deformation mechanism changes from fracturing to crystal plastic processes (see under "mechanisms" below)
Competency	Relative term comparing the resistance of rocks to flow
Failure stress	Stress at which failure occurs
Fracturing	Deformation mechanism by which a rock body or mineral loses coherency
Crystal plasticity	Deformation mechanism that involves breaking of atomic bonds without the material losing coherency
Strength	Stress that a material can support before failure
Ultimate strength	Maximum stress that a material undergoing work softening can support before failure
Work hardening	Condition in which stress necessary to continue deformation experiment increases
Work softening	Condition in which stress necessary to continue deformation experiment decreases
Yield stress	Stress at which permanent strain occurs
Material behavior	
Brittle behavior	Response of a solid material to stress during which the rock loses continuity (cohesion). Brittle behavior reflects the occurrence of brittle deformation mechanisms. It occurs only when stresses exceed a critical value, and thus only occurs after the body has already undergone some elastic and/or plastic behavior. The stress necessary to induce brittle behavior is affected strongly by pressure (stress-sensitive behavior); brittle behavior generally does not occur at high temperatures.
Ductile behavior	A general term for the response of a solid material to stress such that the rock appears to flow mesoscopically like a viscous fluid. In a material that has deformed ductilely, strain is distributed, i.e., strain develops without the formation of mesoscopic discontinuities in the material. Ductile behavior can involve brittle (cataclastic flow) or plastic deformation mechanisms.
Elastic behavior	Response of a solid material to stress such that the material develops an instantaneous, recoverable strain that is linearly proportional to the applied stress. Elastic behavior reflects the occurrence of elastic deformation mechanisms. Rocks can undergo less than a few percent elastic strain before they fail by brittle or plastic mechanisms, and conditions of failure are dependent on pressure and temperature during deformation.
Plastic behavior	Response of a solid material to stress such that when stresses exceed the yield strength of the material, it develops a strain without loss of continuity (i.e., without formation of fractures). Plastic behavior reflects the occurrence of plastic deformation mechanisms, is affected strongly by temperature, and requires time to accumulate (strain rate–sensitive behavior).
Viscous behavior	Response of a liquid material to a stress. As soon as the differential stress becomes greater than zero, a viscous material begins to flow, and the flow rate is proportional to the magnitude of the stress. Viscous deformation takes time to develop.
Deformation mechanisms	
Brittle deformation mechanisms	Mechanisms by which brittle deformation occurs, namely fracture growth and frictional sliding. Fracture growth includes both joint formation and shear rupture formation, and sliding implies faulting. If fracture formation and frictional sliding occur at a grain scale, the resulting deformation is called cataclasis; if cataclasis results in the rock "flowing" like a viscous fluid, then the process is called cataclastic flow.
Elastic deformation mechanisms	Mechanisms by which elastic behavior occurs, namely the bending and stretching, without breaking, of chemical bonds holding atoms or molecules together.
Plastic deformation mechanisms	Mechanisms by which plastic deformation occurs, namely dislocation glide, dislocation creep (glide and climb; including recovery, recrystallization), diffusive mass transfer (grain-boundary diffusion or Coble creep, and diffusion through the grain or Herring-Nabarro creep), grain–boundary sliding/superplasticity.

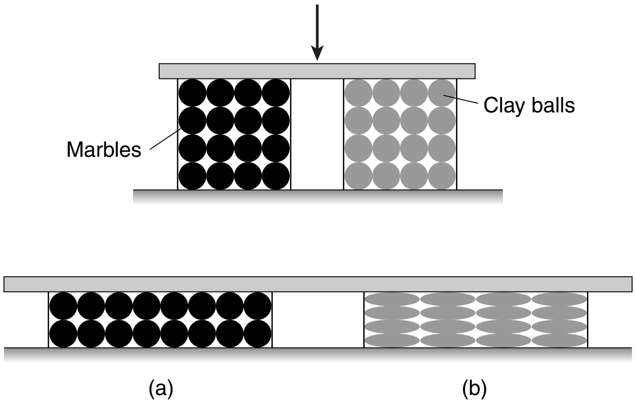


FIGURE 5.19 Deformation experiment with two cubes containing marbles (a), and balls of clay (b), showing ductile strain accumulation by different deformation mechanisms. The finite strain is equal in both cases and the mode of deformation on the scale of the block is distributed (ductile behavior). However, the mechanism by which the deformation occurs is quite different: in (a) frictional sliding of undeformed marbles occurs, while in (b) individual clay balls distort into ellipsoids.

very different. In Figure 5.19a, the rigid spheres slide past one another to accommodate the shape change without distortion of the individual marbles. In Figure 5.19b, the shape change is achieved by changes in the shape of individual clay balls to ellipsoids. In both cases the deformation is not localized, but distributed throughout the block; at the sphere boundaries in Figure 5.19a, and within the spheres in Figure 5.19b. Both experiments are therefore expressions of ductile behavior, although the mechanisms by which deformation occurred are quite different.[20]

Commonly you will encounter the term **brittle-ductile transition** in the literature, but it is not always clear what is meant. For example, seismologists use this term to describe the depth below which non-subduction zone earthquakes no longer occur. If they mean that deformation occurs by a mechanism other than faulting, this usage is incorrect because ductility is not a mechanism. Ductile behavior can occur by faulting if thousands of small cracks take up the strain. Alternatively, ductile behavior may represent deformation in which crystallographic processes are important. We have seen that ductility simply reflects the ability of a material to accumulate significant permanent strain. Clearly, we have to separate behavior (brittle vs. ductile) from mechanism (e.g., fracturing, frictional sliding, crystal plasticity, diffusion, all of which are discussed in later chapters), and in many instances terms such as brittle-ductile transition lead to unnecessary confusion. Instead, a useful solution is to contrast the mechanisms, for example **brittle-plastic transition** or localized versus nonlocalized deformation. Returning to our experiment in Figure 5.19, the marble-filled cube (a) deforms by frictional sliding whereas the clay balls (b) deform plastically. In a rheologic context this is normal stress-dependent and strain rate–dependent behavior, respectively.

In some instances we appear to be faced with a **brittle-ductile paradox.** It is quite common in the field to find folded beds that appear to be closely associated with faults (such as fault-propagation and fault-bend folds, see Chapters 11 and 18). If we assume that faulting and folding occurred simultaneously, then we are left with a situation where both brittle faulting and ductile folding took place at essentially the same level in the Earth. How can we explain this situation and why does it only *appear* to be contradictory? There is no reason to expect that P_c, T, and P_f are sufficiently different over the relatively small volume of rock to account for the simultaneous occurrence of these two behavioral modes of deformation. Strain rate and strain rate gradients, on the other hand, may vary considerably in any given body of rock. Recall the Silly Putty® experiment in which fracturing occurred at high strain rates and ductile flow at lower rates. Similarly, faulting in nature may occur at regions of high strain rates, but some distance away the strain rate may be sufficiently low to give rise to ductile folding. So strain rate gradients are one explanation for the simultaneous occurrence of brittle and ductile structures, which resolves the apparent brittle-ductile paradox.

Competency and strength are two related terms that describe the relationship of rocks to stress. *Strength* is the stress that a material can support before failure.[21] **Competency** is a relative term that compares the resistance of rocks to flow (for example, Figures 5.7 and 5.13). Experiments and general field observations have given us a qualitative *competency scale* for rocks. The competency of sedimentary rocks increases in the order:

rock salt → shale → limestone → greywacke → sandstone → dolomite

---

[20] These mechanisms are frictional sliding and crystal plasticity, respectively.

[21] In the case of work softening, we use the term *ultimate strength* (Figure 5.18).

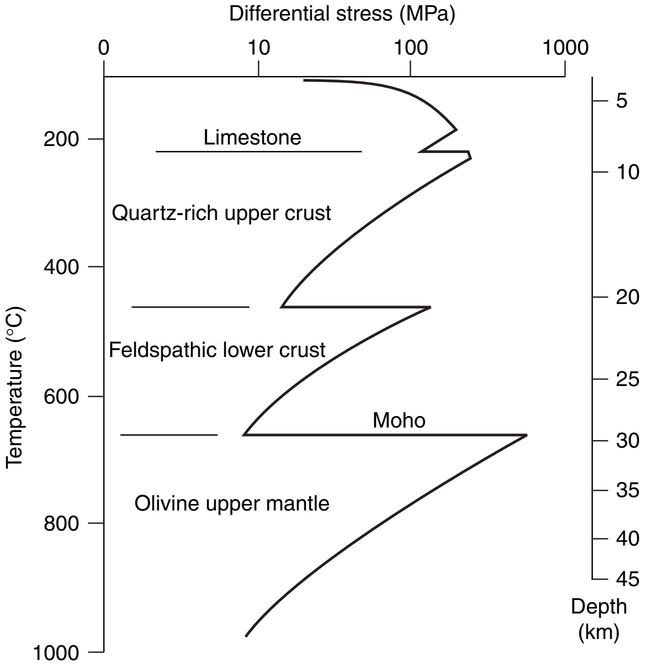


FIGURE 5.20 Rheologic stratification of the lithosphere based on the mechanical properties of characteristic minerals. Computed lithospheric strength (i.e., the differential stress) changes not only as a function of composition, but also as a function of depth (i.e., temperature).

For metamorphic/igneous rocks the order of increasing competency is:

schist → marble → quartzite → gneiss → granite → basalt

Note that competency is not the same as the amount of strain that can accumulate in a body. Therefore ductility contrast between materials should not be used as a synonym for competency contrast.

5.6 CLOSING REMARKS

With this chapter on rheology we conclude Part A on the fundamentals of rock deformation, containing the necessary background to examine the significance of natural deformation structures on all scales, from mountain belts to thin sections. In fact, we can predict broad rheologic properties of the whole earth, given assumptions on mineralogy and temperature. Figure 5.20 shows a composite strength curve for the upper part of the mantle and crust, based on experimental data such as those discussed in this chapter. These "Christmas tree" strength profiles emphasize the role of Earth's compositional stratification and provide reasonable, first-order predictions of rheology for use in numerical models of whole earth dynamics.

Much of the material that we presented in this chapter on rheology takes a basic approach. We examined mainly monomineralic rocks, such as marbles, as opposed to polymineralic rocks, such as granites. Intuitively you will realize that polymineralic rocks require an understanding of the behavior of each of the various constituents, but it is even more complicated than that. Consider the behavior of glass. Glass itself breaks quite easily; however, when glass needles are embedded in resin (epoxy), the **composite material** is unexpectedly strong and resistant to breaking. Glass fiber, consisting of glass and resin, combines the strength of glass with the flexibility of resin. Thus, the behavior of composite materials, and by inference polymineralic rocks, is not simply a matter of knowing the behavior of its constituents, but reflects a complex interplay of properties.

We have chosen to subdivide our discussion of natural deformation in subsequent chapters into brittle and ductile structures. In spite of the sometimes confusing usage of these terms, they do allow a convenient distinction for describing the behavior of natural rocks in the field. Figure 5.21 schematically highlights the mesoscopic aspect of this subdivision, without inferring the mechanisms by which the features form; the subsequent chapters will take care of the latter. This figure also illustrates the broad separation of three types of common geologic structures, faults, folds, and shear zones.

Aside from all the detailed information that deformation structures provide about Earth's history (and that is discussed at length in the subsequent chapters), you should not forget to simply enjoy the sheer beauty and enormity of deformation structures such as faults and folds. This is what attracted many of us to this field of study in the first place.

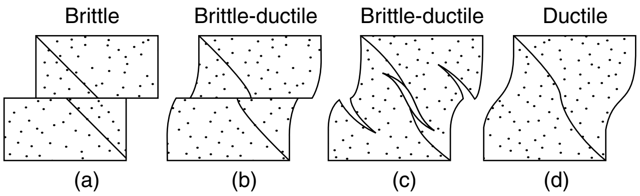


FIGURE 5.21 Brittle (a) to brittle-ductile (b, c) to ductile (d) deformation, reflecting the general subdivision between faults and folds that is used in the subsequent parts of the text.

ADDITIONAL READING

Carmichael, R. S., ed., 1989. *Practical handbook of physical properties of rocks and minerals.* CRC Press: Boca Raton.

Jaeger, J. C., and Cook, N. G. W., 1976. *Fundamentals of rock mechanics.* Chapman and Hall: London.

Kirby, S. H., and Kronenberg, A. K., 1987. Rheology of the lithosphere: selected topics. *Reviews of Geophysics,* 25, 1219–1244.

Poirier, J.-P., 1985. *Creep of crystals. High-temperature deformation processes in metals, ceramics and minerals.* Cambridge University Press: Cambridge.

Ranalli, G., 1995. *Rheology of the Earth* (2nd edition). Allen and Unwin: Boston, 366 pp.

Rutter, E., 1986. On the nomenclature of mode of failure transitions in rocks. *Tectonophysics,* 122, 381–387.