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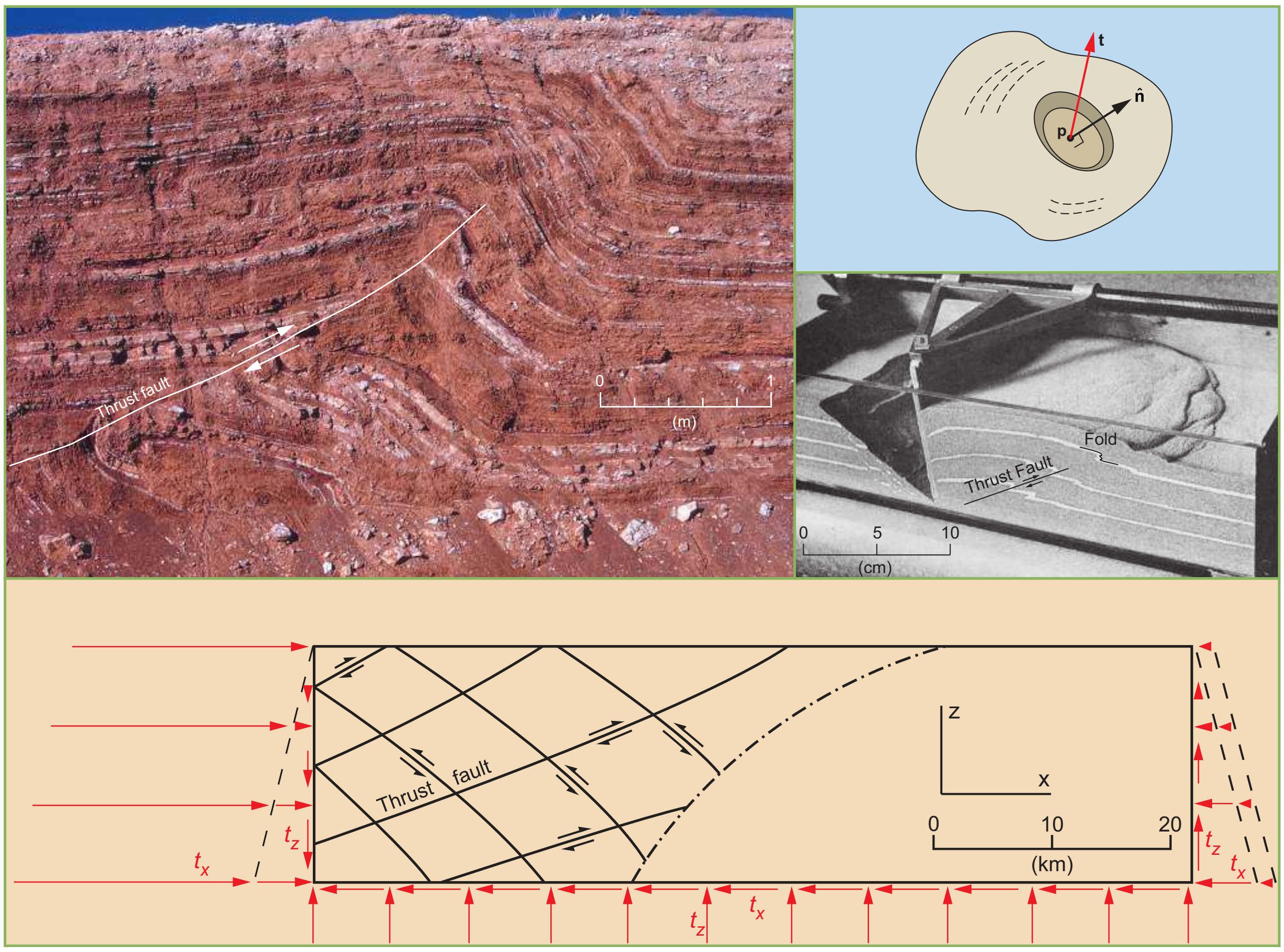
# PART I

Part I (Chapter 1) summarizes the scope of structural geology by addressing the following questions:

- What forces cause deformation of rock in Earth's lithosphere and asthenosphere?
- What are the three major mechanical styles for this deformation?
- What five broad categories of geologic structures result from this deformation?
- What is the methodology advocated in this textbook for analyzing geologic structures?



# Chapter 1

# Scope of Structural Geology

## Introduction

Chapter 1 sets the stage for a quantitative introduction to structural geology. We begin by identifying forces that cause deformation in Earth's lithosphere and asthenosphere. Then, we describe three different styles of deformation, and five broad classes of geologic structures that result from this deformation. To lay out the methodology for studying geologic structures, we introduce what we mean by a complete mechanics and by canonical models of structural geology. Then we examine the roles of physics and mathematics in studying the origins of geologic structures. Finally, we describe applications of structural geology to problems facing our society and the careers that utilize structural geology to solve those problems.

Geologic structures develop when rock deforms from its original configuration into some different configuration. For example, sedimentary rocks typically form in horizontal and tabular layers, but later these layers may tilt into a geologic structure called a *fold* (**Figure 1.1**). To understand the process of folding one must be able to characterize both the initial and the final configurations, identify the material properties of the rock that resisted the deformation, and deduce the forces that caused the deformation. Characterizing the geometry of structures and their progenitors requires data obtained by geologic mapping, and a good understanding of geologic history. Tests carried out in a rock mechanics laboratory measure the material properties of rock. The underlying theory that relates forces to deformation comes from that part of physics called


**Figure 1.1** This fold is located in the Rainbow Basin Natural Area, 13 km north of Barstow in the Mojave Desert of southern California. These sedimentary rocks formed in horizontal layers during the Miocene epoch. Today they tilt about 20° from horizontal. Because the two limbs of the fold are inclined toward each other, this is a syncline.

mechanics, and the language of mechanics is mathematics. Thus, structural geology is a blending of geological knowledge and field techniques, with the results from laboratory tests and the physical principles and mathematical methods of mechanics.

Structural geologists study geologic structures and seek to understand Earth history by placing structures in the proper sequence and dating the tectonic events that generated the structures. They use mechanical principles to understand how and why the structures formed. Structural geologists also help explore for and produce Earth's resources, mitigate geologic hazards, and manage a sustainable environment. As such, structural geologists work in academia, in industry, for governmental agencies, and as private consultants. They work in laboratories, analyze data using computers, and collect geologic data and make maps in the field. In the course of their work structural geologists interface with other earth scientists, with engineers, and with the general public. In these many ways structural geologists contribute to our society.

## 1.1 DEFORMATION OF EARTH'S LITHOSPHERE AND ASTHENOSPHERE

Geophysical data support a primary division of the solid Earth into an inner core, outer core, mantle, asthenosphere, and lithosphere (Figure 1.2). Structural geologists focus their attention on deformation of rock in the two outer shells, which are distinguished on the basis of their mechanical properties. The lithosphere is the stronger upper layer and the asthenosphere is the weaker lower layer. The boundary between the lithosphere and the asthenosphere is associated with a temperature of about 1,300 °C. Under the oceans the base of the lithosphere is at a depth of 50 to 150 km, and under the continents it is at a depth of 50 to 300 km. The base of the asthenosphere is not well defined, but may be as shallow as 200 km, or as deep as 700 km. In any case, the combined thickness of the lithosphere and asthenosphere is a very small fraction of Earth's equatorial radius, which is about 6,380 km.

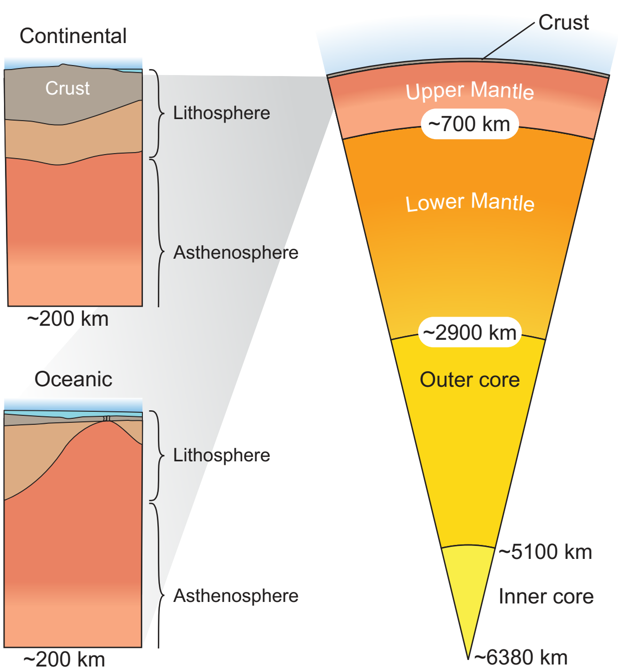



Figure 1.2 Schematic illustration of Earth's primary shells (on the right) including the inner and outer core, lower mantle, upper mantle, and crust. The mantle and crust are classified based on different *composition*. The two outer shells called the asthenosphere and lithosphere (on the left) are classified based on different *strength*. See Press et al. (2004) in Further Reading.

The lithosphere is characterized by *solid mechanical behavior* with localized deformation exemplified by brittle fracturing and faulting in the upper part, and with more distributed deformation exemplified by plastic flow in the lower part. The asthenosphere is characterized by *fluid mechanical behavior* with broadly distributed viscous flow. The change in mechanical behavior from solid-like to fluid-like depends upon the composition of the rock, and also on the temperature and pressure, both of which increase with depth. Increased pressure tends to increase the strength of rock, whereas increased temperature tends to decrease the strength. Within the lithosphere the effect of pressure dominates; hence the apparent strength and rigidity of this outer shell. Below the lithosphere–asthenosphere boundary the effect of temperature dominates; hence the apparent weakness and mobility of this underlying shell. To elucidate these different mechanical behaviors we devote Chapter 4 of this book to *elastic* solid behavior, Chapter 5 to *plastic* solid behavior, and Chapter 6 to *viscous* fluid behavior.

The strong lithosphere composes the plates of plate tectonics (Figure 1.2). The *idealized* concept of plate tectonics is that perfectly rigid plates move laterally, because they are carried along by large-scale convection of the flowing asthenosphere. New material is added to the lithosphere from the upwelling asthenosphere beneath oceanic ridges, and lithosphere is consumed back into the down going asthenosphere along subduction zones. The study of Earth's deformation at global length scales that are relevant to plate tectonics is called geodynamics and textbooks that cover that discipline (see Further Reading at the end of this chapter) overlap some of the topics of structural geology.

If tectonic plates really were perfectly rigid, structural geologists would have very little to study, because only those rocks below the lithosphere–asthenosphere boundary would be deformed. In fact, Earth's lithosphere does deform and this deformation produces a diverse suite of structures including *fractures*, *faults*, *folds*, *fabrics*, and *intrusions*. We devote Chapters 7 through 11 to these categories of structures. They form in many different geologic and tectonic settings, and under broad ranges of temperature, pressure, and rate of deformation. They develop over length scales from millimeters to kilometers and time scales from milliseconds to millions of years. It would be daunting to investigate geologic structures across these immense

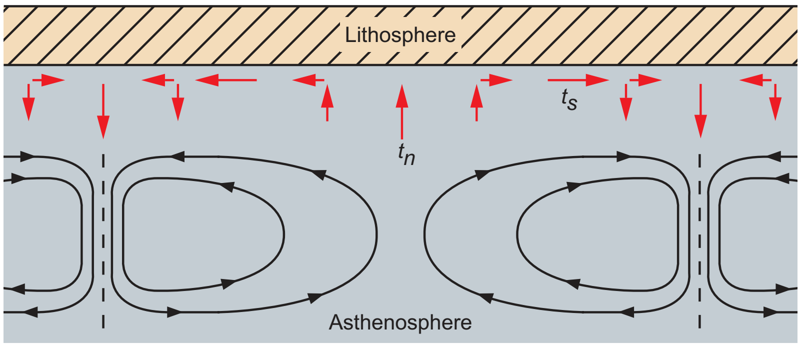



Figure 1.3 Schematic (not to scale) conceptual model of convecting asthenosphere and overlying lithosphere from one of the classic papers on crustal deformation. Closed curves parallel to velocity vectors (black) are flow lines for convection cells; discrete arrows (red) are normal and shear tractions (forces per unit area) acting on the base of lithosphere. Modified from Hafner (1951).

time and length scales, under such a wide range of conditions, and in such disparate rocks without a unifying theory. Continuum mechanics provides the unifying theory, and therefore is given special attention in this textbook.

Earth's two outer shells (Figure 1.2) also may be distinguished on the basis of their chemical composition and mineralogy. The study of Earth's composition in this broad sense is the subject of petrology and geochemistry. The upper shell, referred to as the crust, is made up of various *sedimentary*, *metamorphic*, and *igneous* rocks that are relatively rich in silica, and are composed primarily of the minerals feldspar and quartz. The lower shell, called the mantle, is primarily peridotite, an *igneous* rock that is relatively poor in silica and is composed mostly of the minerals olivine and pyroxene. Due in large part to the different densities of these common minerals, crustal rocks typically are less dense than mantle rocks. The oceanic crust is up to about 10 km thick and the continental crust is about 50 km thick, whereas the mantle is about 1,800 km thick. Thus, the lithosphere typically includes the crust and a portion of the upper mantle, whereas the asthenosphere is a portion of the mantle that is a few to several hundred kilometers thick, below the lithosphere–asthenosphere boundary.

Because the objective of structural geology is to characterize and understand the *deformation* of rock to form geologic structures, the classification of Earth's outer shells based on mechanical properties is more germane than the classification based on composition and mineralogy. However, the physical properties of rock do depend upon composition and mineralogy, so both classifications are useful for structural geologists. Because the lithosphere is more accessible, and it contains the majority of Earth's resources that can be produced, and it is the site of most geologic hazards, *we focus primarily on structures in the lithosphere*.

### 1.1.1 Tectonic Surface Forces: Tractions

Why do Earth's lithosphere and asthenosphere deform? In the context of Newtonian mechanics *forces* cause the deformation of rock in Earth's lithosphere and asthenosphere. In general, these forces are categorized as surface forces and body forces, and both play important roles. We focus on the surface forces here and take up body forces in the next section.

On the base of the lithosphere, surface forces per unit area, called tractions, cause the vertical and lateral motion of the tectonic plates (Figure 1.3). Pressure and viscous drag of the convecting asthenosphere induces these tractions on the base of the lithosphere, and these tractions are, in part, responsible for the geologic structures that are the central topic of structural geology.

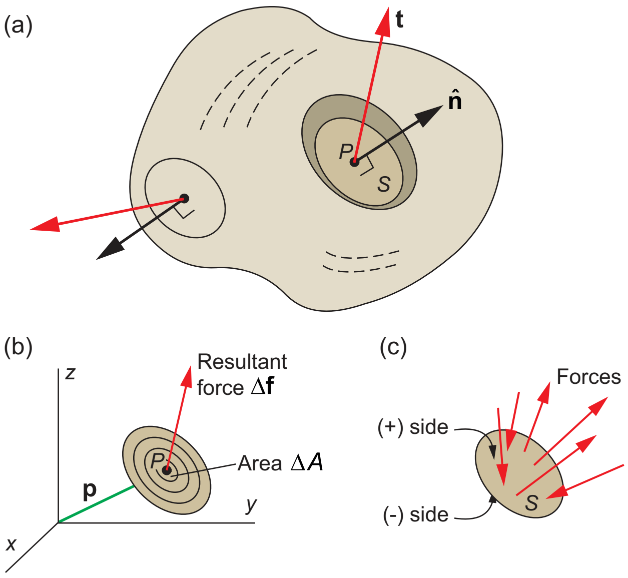



Figure 1.4 The traction vector illustrated. (a) The traction vector **t** acts at the point *P* on the surface *S* with outward unit normal $\hat{\mathbf{n}}$. (b) The point *P* is located by the position vector **p**. On successively smaller patches of *S* the surface area is $\Delta A$ and the resultant force is $\Delta\mathbf{f}$. (c) The mechanical action of the rock on the positive (+) side of the surface *S* acting on the negative (–) side of *S* is represented by distributed forces with resultant $\Delta\mathbf{f}$. The traction vector is defined in equation (1.1).

The traction vector, **t**, is defined as the *surface force per unit area* acting at a point *P* on a surface *S* with a given orientation, either within a rock mass or on its exterior (Figure 1.4a). The surface could be a physical surface, like bedding in sedimentary rock or a compositional boundary in metamorphic rock, or it could be an arbitrarily defined surface. The orientation of the surface is specified using the outward unit normal vector $\hat{\mathbf{n}}$. The point *P* is located by the position vector **p** (Figure 1.4b), using a

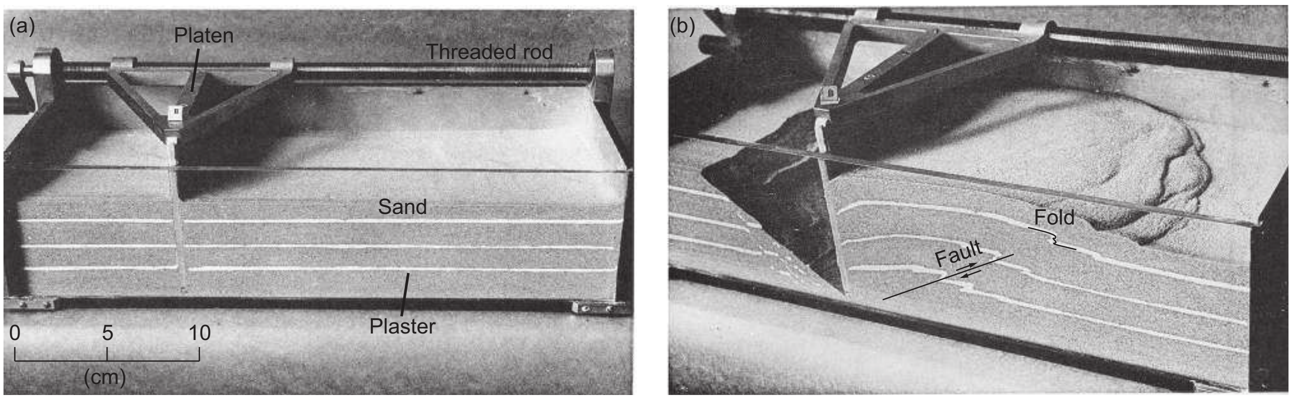



Figure 1.5 Fold and thrust mountain building in the laboratory. (a) Photograph of initial stage of sandbox experiment: layers of sand; interlayers of white plaster. (b) Photograph of final stage after platen moved to the right: asymmetric folds and thrust faults are highlighted by the white plaster. Modified from Hubbert (1951).

Cartesian coordinate system $(x, y, z)$. Note that we use bold letters to represent vectors.

The traction $\mathbf{t}$ represents the mechanical action of the rock on the positive (+) side of the surface $S$, acting on the rock on the negative (–) side of $S$ (Figure 1.4c). This mechanical action is quantified using the forces distributed on $S$, but these could be quite diverse in magnitude and direction if $S$ is large. To focus on the point $P$, we consider some number, $k$, of successively smaller patches of $S$ that all include the point $P$ (Figure 1.4b). For each patch the vector sum (resultant) of the distributed forces is called $\Delta\mathbf{f}$ and the area is called $\Delta A$. The traction is defined as the ratio of resultant force to surface area, in the limit as $k$ goes to an infinite value:

$$\mathbf{t}(\mathbf{p}, \hat{\mathbf{n}}) = \lim_{k\to\infty} [\Delta\mathbf{f}/\Delta A] \tag{1.1}$$

In this limit we insist that the longest dimension of $S$ goes to zero, so the patch goes to a point, not a line. Also, although both $\Delta\mathbf{f}$ and $\Delta A$ go toward zero as the patch goes toward a point, *one of the basic tenets of continuum mechanics is that the ratio, $\Delta\mathbf{f}/\Delta A$, approaches a well-defined limit*.

On the left side of (1.1) the quantities $(\mathbf{p}, \hat{\mathbf{n}})$ are a reminder that the traction vector depends upon the *location*, $\mathbf{p}$, of the point and the *orientation*, $\hat{\mathbf{n}}$, of the surface. At the same point in a rock mass, the traction may differ on surfaces with different orientations passing through that point. Thus, the traction vectors in Figure 1.3 are profoundly different from the velocity vectors in that figure, which have a unique magnitude and direction at a given point. The traction vector at a given point is unique only because we have specified the surface to be the bottom of the lithosphere. If the surface were taken as a vertical plane through that point, instead of the horizontal plane, the traction vector would be different. We explore the traction vector in greater detail in later chapters, because it is one of the most important physical quantities in structural geology.

The tractions acting on the base of the lithosphere (Figure 1.3) induce *stresses* within the lithosphere that contribute to the development of geologic structures. We identify the tractions on the boundary of this deforming rock mass and ask: what are the stresses within the lithosphere and how do these stresses *cause* the rock to deform? This is an example of a boundary-value problem, and much of this textbook is devoted to setting up and evaluating such problems, so there will be many opportunities to study the physical quantity we call stress in later chapters.

As an example of a boundary-value problem, consider the initial and final stages of a classic laboratory experiment at the centimeter scale (Figure 1.5), meant to simulate fold and thrust mountain building. The box is filled with loose, dry sand with thin layers of white plaster that act as markers. A rigid steel platen moves along a threaded rod from left to right as the crank is turned, and the platen pushes the sand layers laterally to form asymmetric folds and thrust faults. The tractions imposed by the platen, and those imposed by the base and sides of the sandbox, induce stresses within the sand that *cause* the folds and faults.

In Figure 1.6 the solution to a boundary-value problem, meant to address faulting during fold and thrust mountain building, is illustrated using the calculated state of stress to define the orientation of *thrust faults*. Tractions on the left and right sides of the model, labeled $t_x$, represent the horizontal tectonic forces that *cause* the faulting. These tractions might be related to the tractions on the base of the lithosphere (Figure 1.3), or they might be due to local disturbances within the lithosphere, but they are greater on the left side than on the right side. On the base of the model the traction component $t_z$ supports the weight of the overlying rock, and the component $t_x$ balances the net horizontal traction applied to the model sides. Given this geometry, the material properties of rock, and the prescribed boundary tractions, the principles of continuum mechanics enable one to calculate the stresses within the rock mass, and deduce the orientation and location of faults. One set of thrust faults dips to the left and curves gently upward, similar to those in the sandbox experiment (Figure 1.5).

Natural examples of fold and thrust structures exist at the scale of entire mountain belts, but also at the outcrop scale (Figure 1.7), where they are easier to record in a photograph. In this example, thinly bedded sedimentary rocks are cut and offset by a small thrust fault, and the offset decreases to zero at

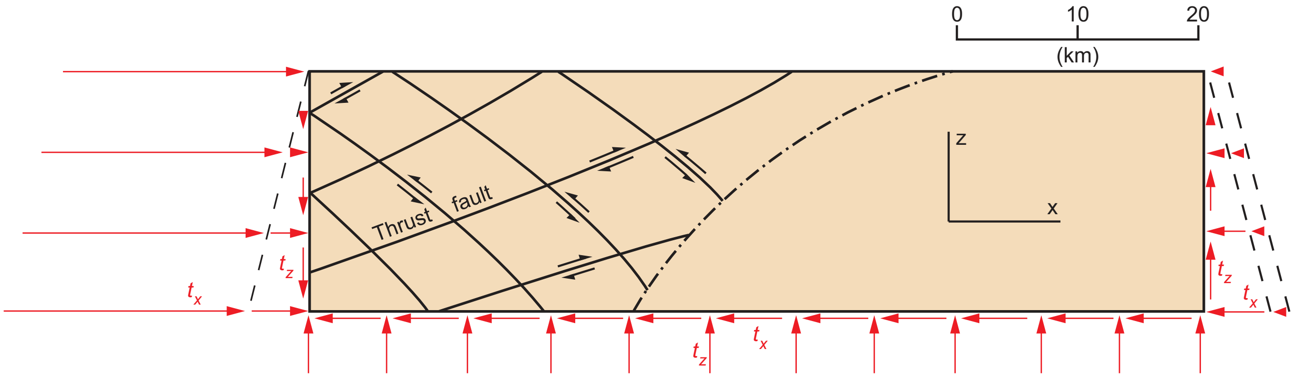



**Figure 1.6** Illustration of a boundary-value problem with applied tractions $t_x$ and $t_z$ on a rectangular block of elastic material. Note 20 km length scale. According to this model stresses within the block cause the thrust faults. Modified from Hafner (1951).

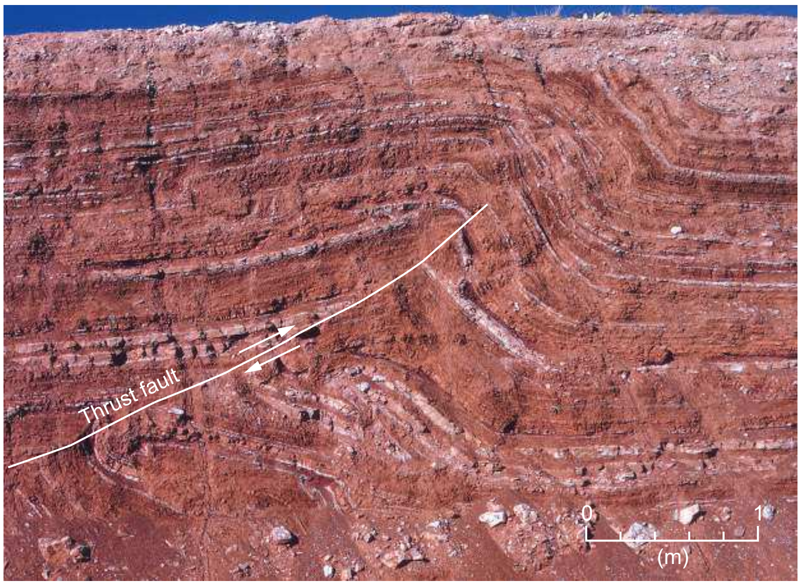



**Figure 1.7** Road cut on State Hwy 276 near Mt Ellsworth in the southern Henry Mountains, UT. Thrust fault offsets and folds the sedimentary strata.

the fault tip. To the right of the fault tip the beds are folded, but continuous. One could define a rectangular domain surrounding this rock mass in its undeformed state, prescribe traction boundary conditions on the sides of that domain, and begin to investigate the formation of these structures in the way illustrated in Figure 1.6. We encounter examples of this procedure later in the textbook.

### 1.1.2 Tectonic Body Forces: Gravity and Buoyancy

Why do Earth's lithosphere and asthenosphere deform? In the previous section we pointed to tectonic surface forces as one cause of this deformation. Here we consider body forces due to *gravity* and *buoyancy* that drive more dense rock downward, and less dense rock upward. Recall that Earth's crust is composed of rock that typically is less dense than the rock in Earth's mantle. Buoyancy, in the broadest sense, explains why crustal rocks sit on top of mantle rocks. For another example, we note that less dense salt moves upward relative to more dense clastic sedimentary rock as salt domes rise in sedimentary basins. Similarly, less dense magma rises to form igneous intrusions in denser host rock.

The gravitational body force, $\mathbf{F}_{\text{grav}}$, acting on a rock mass is:

$$\mathbf{F}_{\text{grav}} = m\mathbf{g} \tag{1.2}$$

In this vector equation, $m$ is the mass of material making up the body, and the gravitational force vector is directed downward, the same direction as the gravitational acceleration, $\mathbf{g}$ (Figure 1.8).

To evaluate the gravitational force we take the gravitation acceleration as uniform and constant over the length and time scales considered, so $\mathbf{g} = \mathbf{g}^*$, and the magnitude of this vector is $g^* = 9.8\ \text{m s}^{-2}$, representative of values in Earth's lithosphere. The Cartesian coordinates are oriented with $z$ positive upward, so the only non-zero component of gravitational acceleration is

$g_z = -g^*$. As an example consider a spherical body of radius $a$, with mass density, $\rho_i$. The mass *inside* the sphere is the volume of the sphere times the mass density: $m = \frac{4}{3}\pi a^3 \rho_i$.

The buoyant force is determined by Archimedes' principle. This famous relationship states that the buoyant force is equal to the weight of the material displaced by the body. Therefore, we write the buoyant force, $\mathbf{F}_{\text{buoy}}$, as:

$$\mathbf{F}_{\text{buoy}} = -m_d \mathbf{g} \tag{1.3}$$

In this vector equation $m_d$ is the displaced mass and $\mathbf{g}$ is the acceleration of gravity. By definition the acceleration vector is directed *downward* and mass is non-negative, so the negative sign on the right side assures that the buoyant force vector is positive, and therefore directed *upward*: it buoys up the less dense body. For example, the mass displaced by a spherical body of radius $a$ is calculated using the volume of the sphere and the mass density, $\rho_o$, of the material *outside* the sphere, such that $m_d = \frac{4}{3}\pi a^3 \rho_o$.

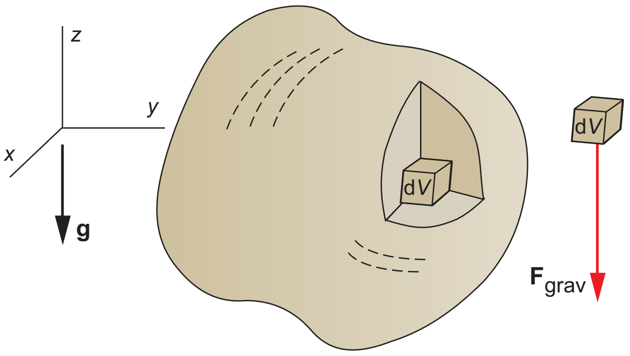



Figure 1.8 The gravitational body force per unit volume, $\mathbf{F}_{\text{grav}}$, is proportional to the mass, $m$, and the gravitational acceleration, $\mathbf{g}$, which defines the direction *down*. The body force acts in the down direction on the small volume element, $dV$.

Adding the gravitational force (1.2) and buoyant force (1.3), and substituting for the respective masses, the resultant body force is:

$$\begin{aligned}\mathbf{F}_{\text{grav}} + \mathbf{F}_{\text{buoy}} &= (m - m_d)\mathbf{g} \\ &= \left\langle 0,\ 0,\ \tfrac{4}{3}\pi a^3(\rho_o - \rho_i)g^* \right\rangle, \text{ spherical body}\end{aligned} \tag{1.4}$$

In the second line we write the vector equation in terms of the components of the resultant body force. The angular brackets contain the three values, separated by commas, which are the *components* of force in the $x$, $y$, and $z$ coordinate directions (Figure 1.8). If the inside density is less than the outside density, the resultant body force is positive (directed upward), the sphere is *buoyant*, and it would tend to rise. If the density inside and outside are equal, the body force is zero, so the sphere is *neutrally buoyant* and would tend to be *stationary*. If the density inside exceeds that outside, the body force is negative, and the sphere would tend to sink. We say "tend to" because the material surrounding the sphere must deform for the sphere to move.

Structures that form due to gravitational and buoyant body forces can be produced in laboratory experiments. Images from such an experiment (Figure 1.9) show the different forms of a buoyant liquid (black oil) rising in a more dense liquid (clear syrup). The two fluids initially fill the rigid box with the less dense fluid on top, and then the box is inverted (stage a). In stages b through d, the interface progressively becomes wavier. In stages e through g, the waves pinch off and rise with a thinning tail. These structures may be analogous to *salt diapirs*, bodies of salt that are buoyant and rise through the overlying sediments. In stages g and h, the tops of the model diapirs flatten against the top of the box and spread laterally. Ultimately, the oil forms a horizontal layer (not shown) at the top of the box.

Structures that form due to gravitational and buoyant body forces also can be studied using boundary-value problems from fluid mechanics. For example, Figure 1.10 illustrates the velocity field for a spherical body of viscous liquid rising in a more dense viscous liquid because of the buoyant force. The velocity

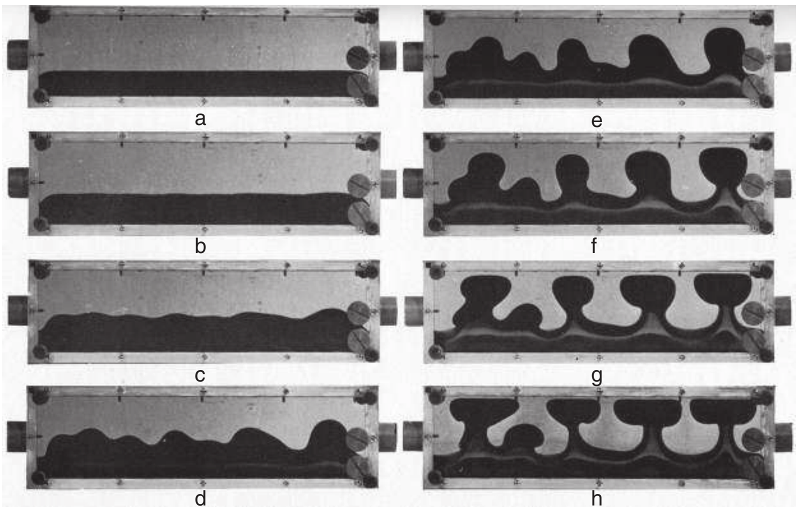



Figure 1.9 Rigid transparent box filled with two fluids. In stages a through h less dense oil (black) rises in more dense syrup (clear) forming diapirs with tails. Modified from Ramberg (1967).

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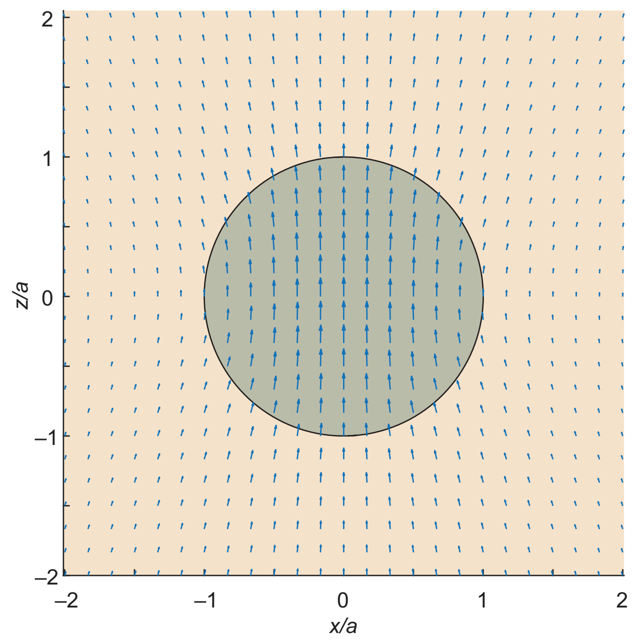


Figure 1.10 Cross section of a sphere of viscous fluid rising in a fluid of different viscosity and greater density. Velocity vectors (arrows) illustrate how the two fluids move as the sphere rises relative to a stationary observer a great distance from the sphere. The equations describing this velocity field are derived in Section 11.6.

field inside and outside the sphere match at the boundary. This solution to the viscous boundary-value problem provides a model for rising salt or magma bodies in a deformable host rock. We explore this solution for the viscous sphere in greater detail in Chapters 2 and 11. In this and many other ways, *gravitational* and *buoyant forces* contribute to deformation of the lithosphere and asthenosphere.

## 1.2 STYLES OF DEFORMATION

How do Earth's lithosphere and asthenosphere deform? Broadly speaking the answer includes three distinctive styles of deformation during which rock *fractures* like a brittle solid, *flows* like a ductile solid, or *flows* like a viscous liquid. These different styles of deformation, or combinations of them, produce the geologic structures that are the subject matter of structural geology. We devote one chapter to each of these styles of deformation, but here we introduce them using evocative pictures that suggest the profound differences between brittle and ductile solids, and between ductile and viscous flow.

### 1.2.1 Brittle

Fractures are the most obvious sign of brittle deformation, and fractures that *opened* as they formed are called veins or joints. Veins are sealed with minerals deposited from groundwater, whereas joints are barren. For example, the *sedimentary rock* shown in Figure 1.11 includes a bed of gray limestone cut by veins that opened and were filled with white calcite. The limestone also is broken by a host of joints with various orientations, and none of these fractures offset the bedding surface, which locally is approximately planar. The joints gap open, and although some of this opening may be due to weathering, some of it records the fact that the joints opened as the limestone was pulled apart in the plane of the bed.

Faults are another sign of brittle deformation. They are distinguished from veins and joints because they *sheared* as they formed. A fault crops out across the center of Figure 1.11 and dips steeply to the left. We identify this structure as a fault because the gray limestone to the left is juxtaposed against a black shale that crops out on the right side of the fault at the same level as the limestone. These strata were deposited at different times, and each was continuous across the plane of the fault before it slipped. Striations, called slickenlines, on a prominent surface of the fault suggest the shearing was directed along the dip of the fault, but the sense of relative motion is not obvious in this image. Did the gray limestone slide up the fault, so it is older than the shale, or did it slide down the fault, so it is younger? These and other questions could be addressed by detailed mapping that identifies the positions of these beds in the stratigraphy.

In Chapter 4 we review data from laboratory tests that help to define the conditions of temperature and pressure under which rocks fracture and fault as they deform. We also describe the strength of rock and its resistance to fracturing and faulting. Test results such as these help structural geologists interpret outcrops with rock broken by veins, joints, and faults.

### 1.2.2 Ductile

The most obvious sign of ductile deformation is pronounced distortion without fracture. For example, the outcrop photograph in Figure 1.12 shows a *metamorphic rock*, called the Moine Schist, made up of highly contorted layers, alternately rich in quartz (white) and mica (gray). Originally this was a sedimentary rock, and these were alternating planar layers of sandstone and shale. Metamorphism has altered the quartz-rich layers, for example, from sand-sized grains of quartz with open pores providing perhaps 20% porosity, to larger interlocked crystals of quartz with near zero porosity. This alteration occurred by plastic deformation and re-crystallization, but the temperature of metamorphism was insufficient to melt the rock. On the other hand the temperature and pressure of metamorphism were sufficient to prevent fracturing as these layers were contorted into very ornate folds.

In Chapter 5 we review data from laboratory tests that help to define the conditions of temperature and pressure under which rocks flow as ductile solids as they deform. We also describe the strength of rock under these conditions, and the physical mechanisms that enable minerals such as quartz to deform without fracturing. These test results help structural geologists interpret outcrops, such as that in Figure 1.12, with folded and sheared rock that did not fracture or fault as it deformed.


Figure 1.11 Three types of brittle fractures in Lower Jurassic limestone beds at Lilstock Beach, Somerset, England. Vertical calcite filled veins (white) and open joints (black) cut the light gray limestone on the left. A fault with striations runs up the middle of the image juxtaposing the gray limestone with the black shale on the right. Scale is approximate in the foreground. See Engelder and Peacock (2001) for detailed maps, outcrop photos, and structural interpretations of this area. Google Earth file: Figure 1.11 Lilstock Beach England fractures.kmz. UTM: 30 U 485890.63 m E, 5672212.54 m N (see Box 1.1 for an explanation of UTM).

### 1.2.3 Viscous

Fissure eruptions near the summit of Kilauea Volcano on the island of Hawaii produce ample evidence for viscous deformation: molten rock behaves like a viscous liquid. A nighttime image (Figure 1.13) shows a curtain of fire erupting from a fissure in the background and lava flowing over a cliff that surrounds a crater in the foreground. There can be no doubt that the mechanical behavior of lava is that of a liquid, but its slow progress down the slope of the volcano, compared to water flowing down a similar slope, suggests lava has a greater resistance to flow. Resistance to flow is measured by the material property called *viscosity*, and the viscosity of basaltic lava is as much as six orders of magnitude greater than that of water.

In Chapter 6 we review laboratory tests that measure viscosity and melting as a function of temperature and chemistry. Because rock is a multi-component system, usually consisting of several different minerals, it has a range of melting temperatures. At the solidus temperature all constituents are glass or crystals, and at the liquidus temperature all constituents have melted. For basaltic lava (Figure 1.13), the solidus is about 1,000 °C and the liquidus is about 1,200 °C.

## 1.3 GEOLOGIC STRUCTURES

What are the products of deformation in Earth's lithosphere? We organize the products of deformation into five different categories of geologic structures: fractures, faults, folds, fabrics, and intrusions. The tectonic surface forces (tractions) and body forces (gravity and buoyancy) *cause* brittle, ductile, and viscous deformation, and the ensuing relative motions produce these structures. We devote a chapter to each category, but introduce them here with a few images.

### 1.3.1 Fractures

*Opening* fractures are the most common structures that form by brittle deformation in Earth's lithosphere. They occur within individual mineral grains at scales of microns to millimeters, and in outcrops such as the Cedar Mesa Sandstone (Figure 1.14) at scales of meters to kilometers. In this example the fractures are vertical and are organized into two nearly orthogonal sets, labeled A and B, based on their orientation. They are regularly spaced, but appear to be confined to the mesa-forming sandstone unit.