CHAPTER 7: DEFORMATION OF ROCKS – STRUCTURAL GEOLOGY

In this chapter you will learn about some of the structures that are produced when rocks are deformed by stresses in the crust. For a more detailed treatment of the deformation of rocks and its causes I recommend Park (2012)

(Note: Terms in red and italics appear as entries in the companion glossary.)

7.1 BRITTLE AND DUCTILE DEFORMATION

As we have seen in Chapter 6 on metamorphic rocks, stress will deform crustal materials, which is something that can happen at any temperature or pressure. Deformation can be either brittle or ductile, depending on temperature and pressure, how fast stress is applied, and the mechanical properties of the rocks (Fig. 7.1). *Brittle deformation* is when rocks fail as rigid solids (Fig. 7.1A). The rocks will break, rather than bend, under these conditions to produce fractures. Brittle deformation occurs along discrete planes in the rock instead of involving the rock body as a whole. Later, fractures can be places where material is removed as a result of rock dissolution, or minerals can grow in the open fracture spaces. Some fractures develop displacements as rock bodies on opposite sides move relative to each other. If this happens, the fracture becomes a *fault*. It takes enormous amounts of stress to generate fractures in some rock formations. When stresses build in the rock, it will not only create a fracture, but tension may create openings in the rock along the fracture. These features are called *tension gashes*, and their orientation tells us how shear stress was oriented on opposite sides of the fracture, or what we call the *sense of shear* (Fig. 7.2). Tension gashes open with an orientation pointing into the direction in which compression was applied. They give us a sense of how displacement would occur if the fracture zone were to become a fault.

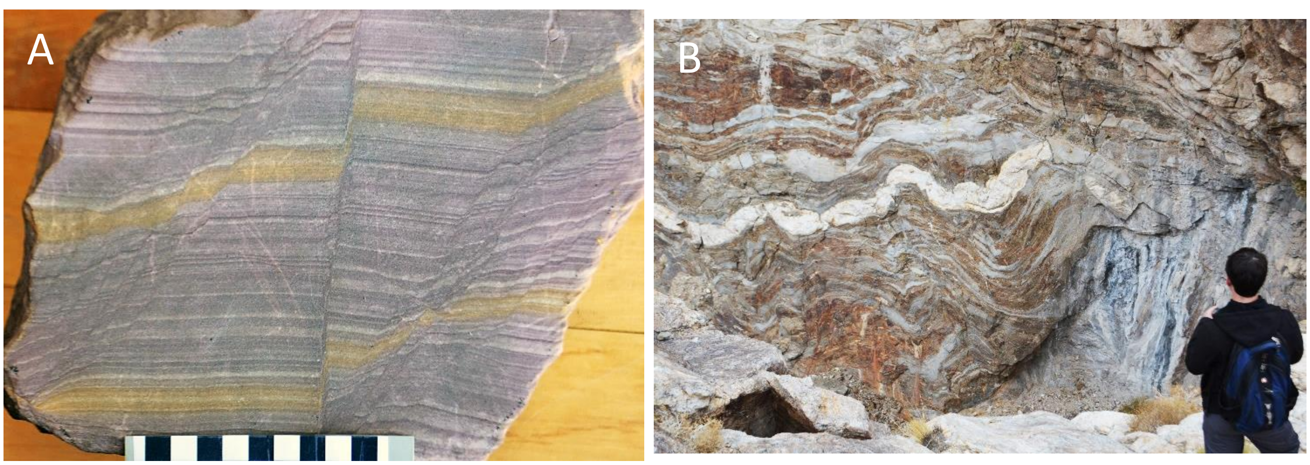

Figure 7.1 *– Brittle vs. ductile deformation. A) Brittle deformation in the form of fractures and small faults (discrete planes) in a sandstone sample. Scale in cm. B) Ductile deformation in the Crystal Spring Formation in Chloride Canyon in Death Valley National Park, California.*

Ductile deformation is when a rock deforms plastically like silly putty (Fig. 7.1B). In this case, features in the rock can change shape without developing visible fractures, and the whole material is involved in deformation. It is possible for grains to move relative to each other or rotate, or mineral grains can dissolve and recrystallize. We have already seen evidence of ductile deformation when we discussed the stretching and flattening of features such as pebbles in a

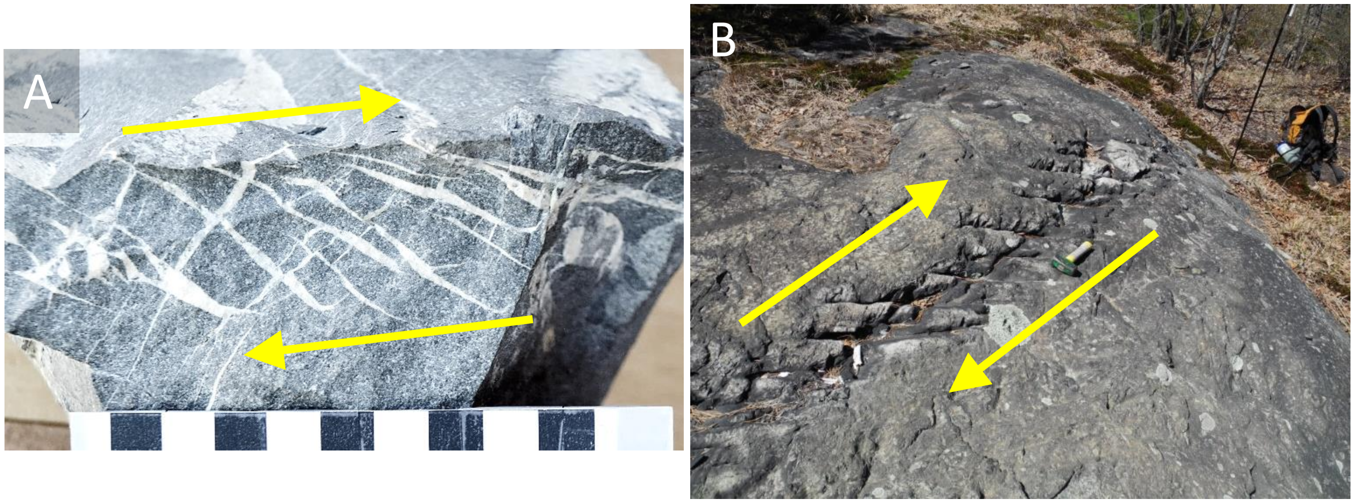

Figure 7.2 *– Tension gashes form as a result of tension along a plane on which shear stress is applied. A) Tension gashes in limestone from Bethlehem, Pennsylvania. The fractures have been filled by calcite cement (white) derived from limestone dissolved in another area. B) A long shear in igneous rock of the Fells that has associated tension gashes. The arrows give the sense of shear.*

conglomerate (see Fig. 6.2B). In some situations, it is possible to determine the amount of ductile deformation, or strain, a rock has undergone by recording the shapes of objects that had some known shape prior to deformation (Fig. 7.3). These strain markers are often circular or spherical objects that are now flattened or sheared to an oval or ovoid shape, or they can be deformed fossils that had a known starting shape that can be seen in non-deformed rocks.

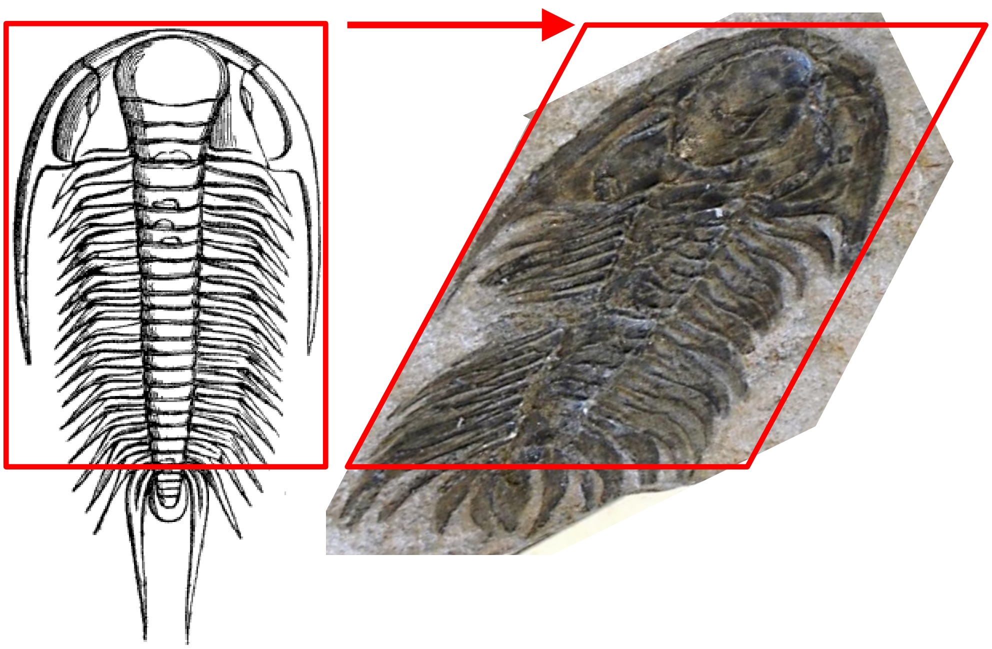

Figure 7.3 *– Sometimes objects in a rock had a known shape prior to deformation and can be used to determine the amount of strain, or shape change, the rock has experienced. This includes fossils of known original shape like the trilobite shown here (Paradoxides bohemicus). (left) Sketch of the undeformed fossil (Figuier, 1863). (right) Sample of the same fossil species, deformed. The arrow and boxes indicate a potential type of deformation, called simple shear, that might be used to explain how the rock was deformed from its original shape, or the amount of strain the rock has experienced. A shortcoming of this reconstruction is that it is 2-dimensional and does not account for how the fossil may have deformed perpendicular to the page.*

7.2 RECOGNIZING FAULTS

When rocks fail along a fracture plane and opposite sides are displaced relative to each other, the plane becomes a *fault*. Faults are shown on geologic maps as lines heavier than the regular contact lines between geologic units (Fig. 7.4). In the field, faults can be recognized in several ways. Faults are frequently not single planes along which displacement occurred but instead are zones in which the rock is heavily sheared by a network of smaller faults and fractures that together make up a *fault zone*. This is a messier situation than most people imagine but should be expected where rocks of different hardness are under stress and are torn apart along surfaces that are seldom smooth or flat. Unfortunately, we often don't get to have a close examination of fault zones in outcrops. Highly fractured rock can be heavily weathered, and is therefore easily eroded, especially in an area like the Fells that was once glaciated. This results in a ravine cut into the land surface because the fault becomes an area of low resistance to erosion (Fig. 7.5). Both sides of the ravine will often have closely-spaced fractures, sometimes filled by veins, that parallel the ravine, and the floor of the ravine may be covered by angular rock debris. Perhaps the best way to demonstrate a fault is to map out the rocks on both sides of the suspected fault or fault zone and see if features on opposite sides are displaced. One of the easiest ways of doing this is to map out the contacts between rock formations, which will be displaced where they are cut by a fault. Dikes are very important in this regard. If they are older than the fault, the dikes will be broken and displaced (Fig. 7.4). Of course, the opposite scenario, in which dikes are younger than the fault, is also interesting because it gives us the relative ages of dikes: those cut by the fault are an older set of dikes, and those that cut the fault are a younger set of dikes. In Chapter 8 you will learn about objects like this as crosscutting features.

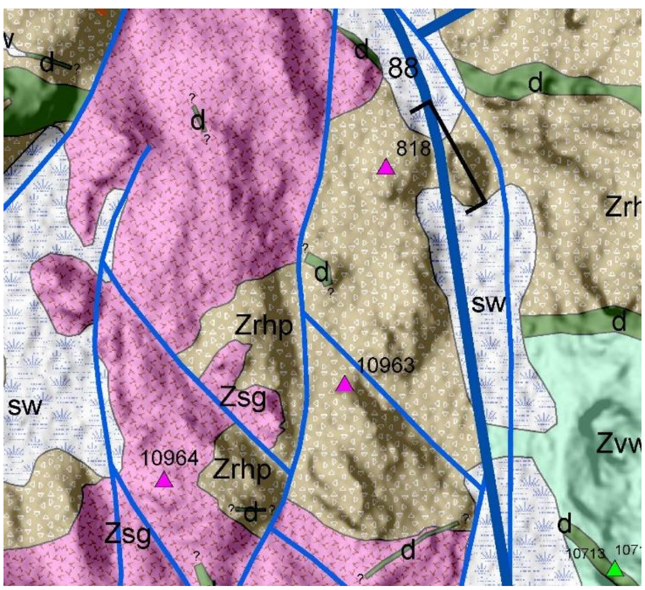

Figure 7.4 *– Faults on geologic maps are recognized by the fact that they displace the contacts of units on opposite sides of the fault. Faults (blue lines) displace the contacts of plutonic igneous rocks (Zrhp, Zsg), metasandstone (Zvw), and some dolerite dikes (d) southeast of South Reservoir in Medford. Note how the major fault (heavy blue line) occurs in a low area with a swamp (sw). The map is about 250 m across.*

Figure 7.5 *– In the field, faults may be indicated by topography and other surface features, such as at this site near the Fellsway East where there is a small, narrow ravine (yellow line) floored and surrounded by shattered rock debris. Alignment of faults with topography can also be seen on the map in Figure 7.4.*

In order to alleviate any concerns that you may have, it should be emphasized that all the faults in the Fells, ***as far as we know***, are dormant or inactive. The Earth and Ocean Sciences Department at Tufts occasionally gets phone calls from concerned citizens who have heard about the North Border Fault that appears on geologic maps south of the Fells and runs through the towns of Medford (along Rt. 16 from Boston Avenue through downtown) and Malden (along Pleasant Street from the Fellsway through downtown). We must assure people that this fault is ancient and is no longer active. However, as a geologist in the Boston area, it would be pretty cool to have an active fault to study.

7.3 TYPES OF FAULTS

Faults are classified as dip-slip or strike-slip faults depending on the relative displacement of rock on opposite sides of the fault. Along *dip-slip faults*, rocks on opposite sides of the fault have moved vertically relative to each other (Fig. 7.6), while along strike-slip faults, rocks on opposite sides have moved horizontally relative to each other. Unless a fault plane is vertical, it will have a dip of less than 90°. All non-vertical faults have one rock body above the fault called the *headwall* and another rock body beneath the fault called the *footwall*. If the headwall moves down relative to the footwall, this type of dip-slip fault is called a *normal fault*. If the opposite is true, and the headwall moves upward on the footwall, this dip-slip fault is called a *reverse fault*. If a reverse fault has a low dip angle (under 30°) it is further classified as a *thrust fault*. These fault types allow us to determine whether there has been tension or compression in the crust. Normal faults result from tension in the crust that pulls rocks apart and creates extension, while reverse and thrust faults result from the crust being compressed and shortened.

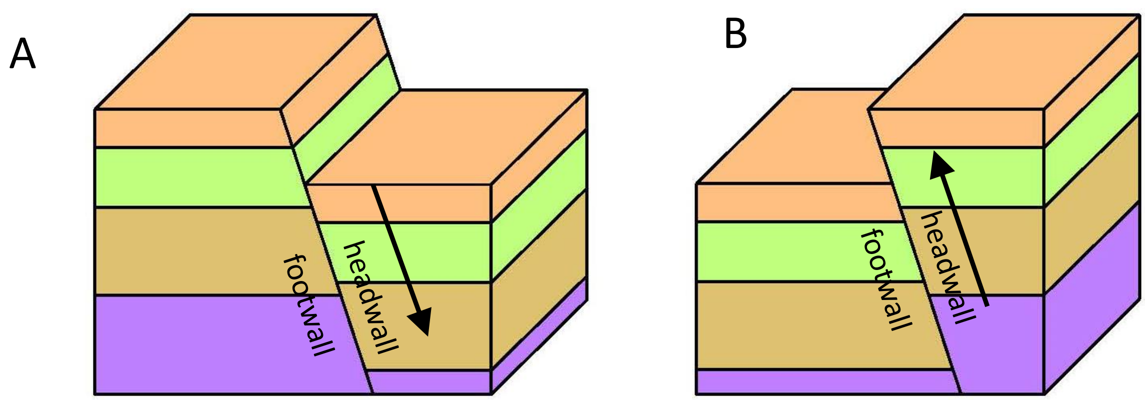

Figure 7.6 *– Dip-slip faults with vertical displacement of rocks on opposite sides of the faults. The arrows show the relative displacement of the headwalls of the faults. A) Normal fault with the headwall moving down the footwall; B) Reverse fault with the headwall moving up the footwall.*

Strike-slip faults can also be broken down into two separate types (Fig. 7.7). If rock on one side of a fault, when viewed from across the fault, appears to move to the right, it is known as a *right-lateral (or dextral) strike-slip fault*. The San Andreas Fault in California is a good example of a right-lateral strike slip fault zone. If rock on one side of a fault, when viewed from across the fault, appears to move to the left, it is known as a *left-lateral (or sinistral) strike-slip fault*. Fault classification would be simple, if it weren't for the fact that horizontal and vertical displacements are often combined. When these oblique motions occur, a fault will take on the name of both types of movement. For example, there can be a normal or reverse left-lateral strike-slip fault, or a normal or reverse right-lateral strike-slip fault. Figuring all this out from field relationships is difficult without seeing displacement of multiple features that have different orientations.

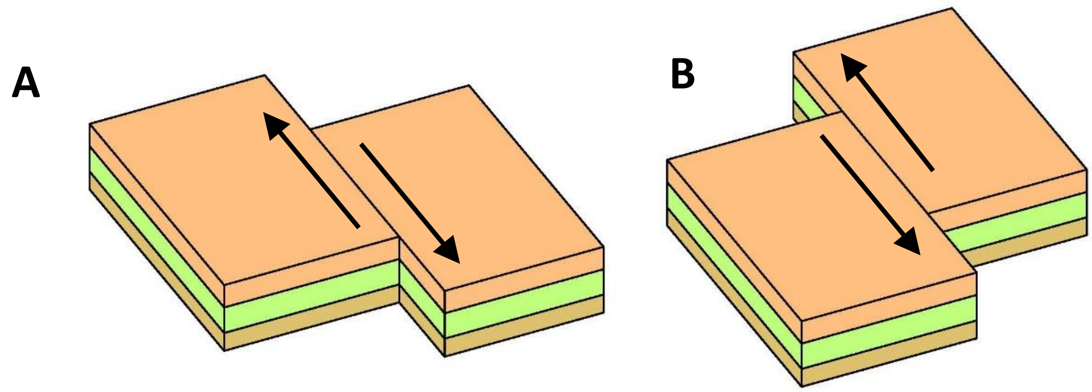

__Figure 7.7__ – Strike-slip faults with horizontal relative movement of rocks on opposite sides of the faults. Arrows show relative displacements. A) Right-lateral strike slip fault. B) Left-lateral strike slip fault. Faults are almost never purely one type of movement but rather are combinations of a vertical (either normal or reverse) and lateral (either right- or left-) movements.

7.4 DETERMINING FAULT DISPLACEMENT DIRECTION

Geologists usually want to determine the direction, or sense, of displacement and the amount of displacement on a fault. This can sometimes be determined by the map pattern of displaced objects. Although this may look relatively straightforward from the diagrams in Figures 7.6 and 7.7, it can be very difficult, depending on the orientation of displaced planar objects and whether faults have a combination of motions (up or down in addition to horizontal displacement). The map pattern itself may be misleading. For example, an object may appear to be displaced horizontally on a map, when in reality there was also dip-slip movement on the fault that isn't apparent in the map pattern (Fig. 7.8). It is also possible to produce a map pattern that appears to show strike-slip movement with just dip-slip movement followed by erosion.

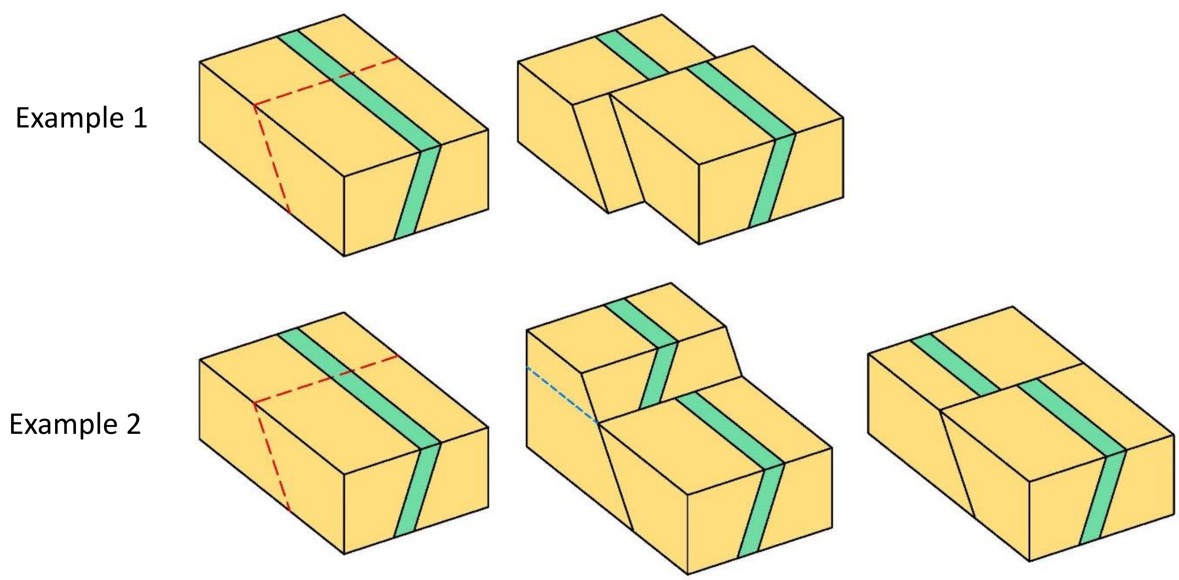

__Figure7.8__ – It can be difficult to tell the true sense of displacement on a fault by simply looking at a map pattern even when the fault plane has a well-known orientation. In Example 1, a left-lateral strike-slip fault with purely horizontal displacement at the position of the dashed red line in the block on the left displaces a dipping dike, producing the map pattern shown on the displaced blocks on the right. In Example 2, displacement of the dike at the red line is along a normal fault with purely vertical motion. After fault displacement, the higher area toward the back is eroded down to the level of the front (dashed blue line) and produces the same map pattern shown in Example 1. Thus, there are at least two ways to get the same map pattern! Many more possibilities exist when you consider combinations of strike-slip and dip-slip motions.

In order to determine the true displacement direction on a fault, geologists rely on fault structures. Displacement of rock surfaces along fault planes produces ground-up rock debris, and minerals often grow along fractures while displacement is occurring. As mineral growth begins in the open spaces of a fault zone, the fault zone debris and recent mineral growth can be broken by continued displacement to create shiny, grooved surfaces called *slickensides*. Slickenside surfaces often have steps as a result of the pulling apart of partly cemented fault material and can be used to indicate the direction of fault movement (Fig. 7.9, see also Fig. 1.8 in Chapter 1).

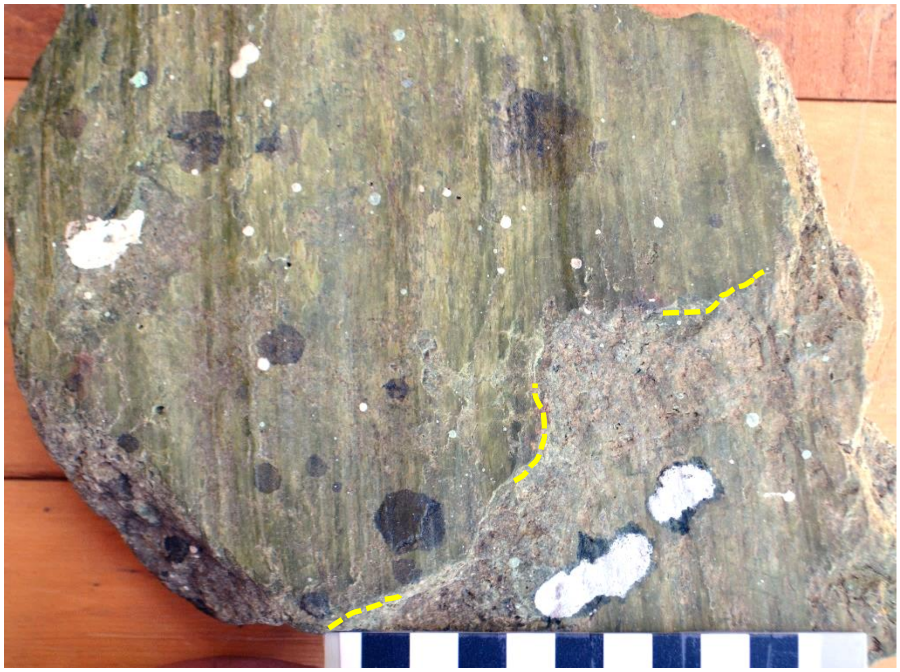

Figure 7.9 *– Slickensides on a sample of an ancient fault plane with epidote mineralization of the grooved surface (see also Fig. 1.8 in Chapter 1). The grooves on the surface allow us to determine the trend of movement on the fault, which is parallel to the sides of the image in this view. Steps on the surface are marked by the yellow dashed lines, marking a break leading down to a lower set of slickensides in the lower right. The steps indicate the relative motion of rocks at the surface. The rock above the slickensides was moving toward the bottom of the view relative to the slickensided surface. Otherwise, the steps would lock up and break by movement on the slickensides toward the top of the view.*

7.3 SIGNIFICANCE OF FAULTS AND FRACTURES

The recognition of faults is not only important to understanding the geologic history of an area, but they also tell us something about land surface stability and whether there has been any recent movement of the crust that could be associated with earthquakes. *Earthquakes* are generated by the buildup of stress in the crust that eventually causes rocks to fail abruptly and move relative to each other. Because earthquakes are generated by this movement, they are always associated with faults. When failure occurs along a fault, it releases large amounts of elastic energy, which travels through the crust as shock waves. Locating active faults is an important part of understanding earthquake hazards. Luckily, as far as we know, there are no active faults in the immediate Boston area. However, the northeastern U.S. is by no means immune to earthquakes (NESEC, 2021). The active fault closest to Boston is to the north near Cape Ann, where an earthquake in 1755 of magnitude 6 on the Richter Scale rocked the northeastern United States and maritime Canada, causing significant damage to colonial structures in eastern Massachusetts and Nova Scotia (Ebel, 2019).

An additional significance of faults and their associated fractures is that they produce openings in the subsurface that allow groundwater to move. In fact, some of the best-producing water wells are in fault zones where there is high permeability in the subsurface due to fractured rock. In some places along old faults and fractures, groundwater has chemically precipitated minerals, especially quartz, epidote, iron oxide/hydroxide and calcite, that today show up as thin fracture fillings called *veins* (Fig. 7.10). Veins look like miniature dikes, but they are formed by chemical precipitation from a water solution and not by intrusion of magma (see also Fig. 1.9C in Chapter 1).

Faults and fractures are interpreted for the information they give about stress. The orientations of faults and fractures are not random but instead follow patterns. They are the product of the orientation of stresses in the crust at the time in the past when the fractures were created, or when a fault was active. Fractures and faults provide pathways for magmas. Dikes are essentially filled-in fractures where tension in the area and pressure from the magma widened a fracture and allowed magma to squeeze through the subsurface. It is common for sets of dikes to run parallel to each other since the original fractures were parallel. Dikes that are parallel to each other were likely responding to the same stress field that opened a particular set of fractures, and the parallel dikes may be the same age. Amazingly, some small dikes (less than a meter wide) are very continuous and can be traced for kilometers.

Figure 7.10 *– Veins, which look like thin mineral dikes, are produced by the precipitation of minerals from water solutions in fracture spaces. Shown here are quartz veins filling both displacements (small faults) that cut across metamorphic rock layers and spaces parallel to layering. Image taken at Bald Head, Maine.*

7.5 FOLDING AND FOLD TYPES

The folding of once flat layers in rocks is the result of compression, or shortening, of Earth’s crust by large-scale forces. Fold patterns are most easily recognized in sedimentary rocks where bedding planes serve as markers of once-flat surfaces. Folds are classified into two general types, *anticlines* and *synclines* (Fig. 7.11). Folds can occur as a part of metamorphic events, or rocks can be folded with very little evidence of metamorphism. In some areas with active compression, folds start to form in sediment that hasn’t yet been cemented. When rocks are brittle, folding tends to produce many fractures. In rocks that are ductile, there can be squeezing of material into the peaks, or noses, of folds. Slatey cleavage (Fig. 7.12) may develop as very fine mica starts to form during metamorphism. The mica tends to grow perpendicular to compressive stresses, and the rock will more easily break in this direction, parallel to the cleavage planes in mica flakes. This may involve substantial material dissolving, and the development of highly parallel break planes on a large scale called rock cleavage. Previously, we discussed folds in banded felsite or obsidian created by mixing and folding of compositional layers in liquid sialic lava flows (Chapter 4). Flow banding is not caused by regional crustal compression like large-scale folds that effect whole rock formations or occur across mountain ranges.

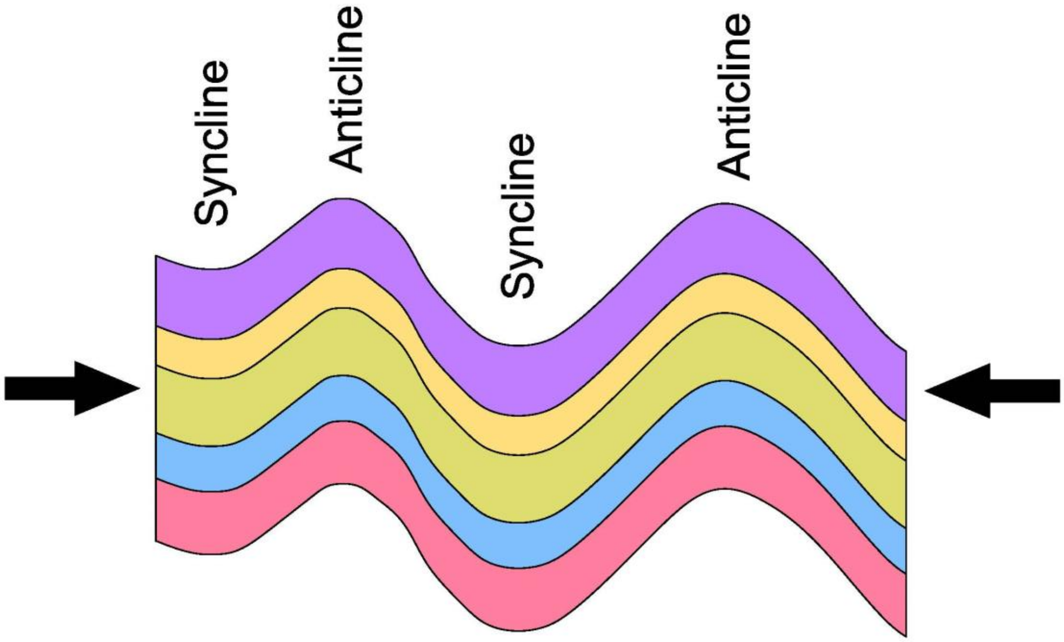

Figure 7.11 *– Folds in an originally horizontal stack of sedimentary beds in cross section. As a result of horizontal compression (arrows), beds are pushed into successive anticlines and synclines.*

Figure 7.12*– Outcrop of slate in the Martinsburg Formation along Rt. 17 in southeastern New York State with a syncline (white line) in the middle of the image. The syncline is tipped over so that the limb on the right is steeper. Cleavage (red) has developed parallel to the axis or center line of the fold (axial planar cleavage) as a result of compression (squeezing in direction of yellow arrows) perpendicular to the cleavage direction.*

Across regions where rocks have been folded, layers that are resistant to erosion tend to remain higher than non-resistant layers. This makes the fold pattern in the rocks easily discernable on satellite images (Fig. 7.13). Many mountain belts have a zone of folded rocks along with low-angle reverse faults called a fold and thrust belt.

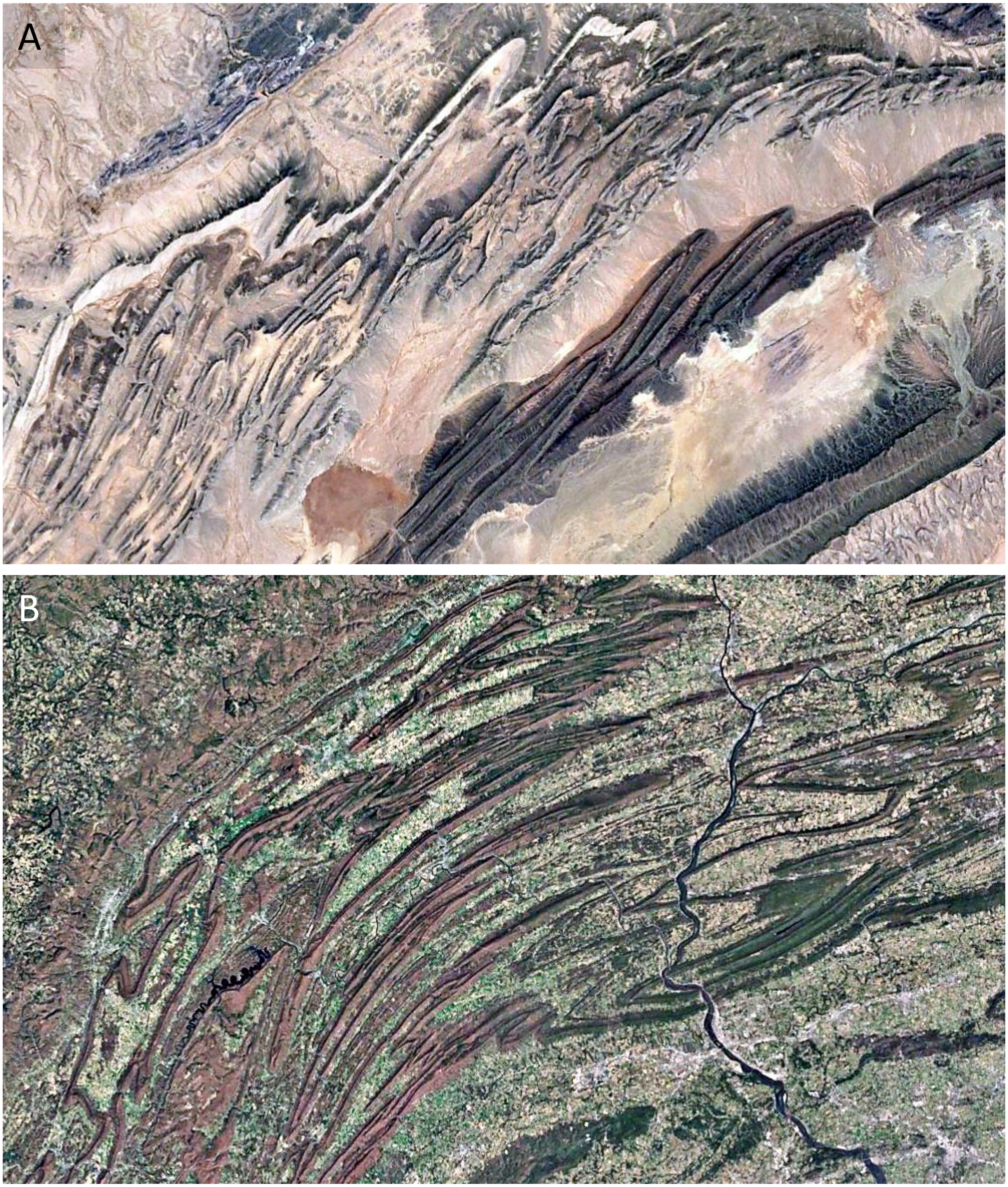

***Figure 7.13** – Google Earth satellite images showing areas of folded sedimentary rocks with patterns of repeated anticlines and synclines. A) The Anti-Atlas Mountains in Morocco. The desert environment of North Africa makes the structure of the rocks easy to see. B) The folded Appalachians in central Pennsylvania. On the right side of the mage is the Susquehanna River cutting across the fold belt from north to south.*

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