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Title of the lecture

Brittle Deformation

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Introduction

Drop a glass on a tile floor and watch as it breaks into dozens of pieces; you have just witnessed an example of brittle deformation! Because you've probably broken a glass or two (or a plate or a vase) and have seen cracked buildings, sidewalks, or roads, you already have an intuitive feel for what brittle deformation is all about. In the upper crust of the Earth, roughly 10 km in depth, rocks primarily undergo brittle deformation, creating a myriad of geologic structures. To understand why these structures, exist, how they form in rocks of the crust, and what they tell us about the conditions of deformation, we must first learn why and how brittle deformation takes place in materials in general. Our purpose is to introduce the basic terminology used to describe brittle deformation, to explain the processes by which brittle deformation takes place, and to describe the physical conditions that lead to brittle deformation.

Vocabulary of brittle deformation

Research in the last few decades has changed the way geologists think about brittle deformation and, as a consequence, the vocabulary of brittle deformation has evolved. Summarizes definitions of terms that we use in discussing brittle deformation, and includes additional terms that you may come across when reading articles on this topic. Remember that some of the brittle deformation vocabulary is controversial, and you will find that not all geologists agree on the definitions that we provide. Brittle deformation is simply the permanent change that occurs in a solid material due to the growth of fractures and/or due to sliding on fractures once they have formed. By this definition, a fracture is any surface of discontinuity, meaning a surface across which the material is no longer bonded. If a fracture fills with minerals precipitated out of a hydrous solution, it is a vein, and if it fills with (igneous or sedimentary) rock originating from elsewhere, it is a dike. A joint is a natural fracture in rock across which there is no measurable shear displacement. Because of the lack of shear involved in joint formation, joints can also be called cracks or tensile fractures. Shear fractures, in contrast, are mesoscopic fractures across which there has been displacement. Sometimes geologists use the term "shear fracture" instead of "fault" when they wish to imply that the amount of shear displacement on the fracture is relatively small, and that the shear displacement accompanied the formation of the fracture in once intact rock. In a broad sense, a fault is a surface or zone on which there has been measurable displacement. In a narrower sense, geologists restrict use of the term fault to a fracture surface on which there has been sliding. When using this narrow definition of fault, we apply the term fault zone to refer either to a band of finite width across which the displacement is partitioned among many smaller faults, or to the zone of rock bordering the fault that has fractured during faulting. We apply the term shear zone to a band of finite width in which the ductile shear strain is significantly greater than in the surrounding rock. Movement in shear zones can be the consequence of cataclasis (distributed fracturing, crushing, and frictional sliding of grains of rock or rock fragments). Regardless of type, a fracture does not extend infinitely in all directions (Figure 1). Some fractures intersect the surface of rock, whereas others terminate within the body. The line representing the intersection of the fracture with the surface of a rock body is the fracture trace, and the line that separates the region of the rock which has fractured from the region that has not fractured is the fracture front. The point at which the fracture trace terminates on the surface of the rock is the fracture tip. In three dimensions, some fractures have irregular surfaces whereas others have geometries that roughly resemble coins or blades.

- **Brittle deformation**: The permanent change that occurs in a solid material due to the growth of fractures and/ or due to sliding on fractures. Brittle deformation only occurs when stresses exceed a critical value, and thus only after a rock has already undergone some elastic and/or plastic behavior.
- **Cataclasis:** A deformation process that involves distributed fracturing, crushing, and frictional sliding of grains or of rock fragments.
- **Crack:** A fracture whose displacement does not involve shear displacement (i.e., a joint).
- **Fracture**: A general term for a surface in a material across which there has been loss of continuity and, therefore, strength. Fractures range in size from grain-scale to continent-scale.
- **Microfracture**: A very small fracture of any type. Microfractures range in size from the dimensions of a single grain to the dimensions of a thin section.
- **Shear zone**: A region of finite width in which ductile shear strain is significantly greater than in the surrounding rock. Movement in shear zones is a consequence of ductile deformation mechanisms (cataclasis, crystal plasticity, diffusion)
- **Vein**: A fracture filled with minerals precipitated from a water solution.

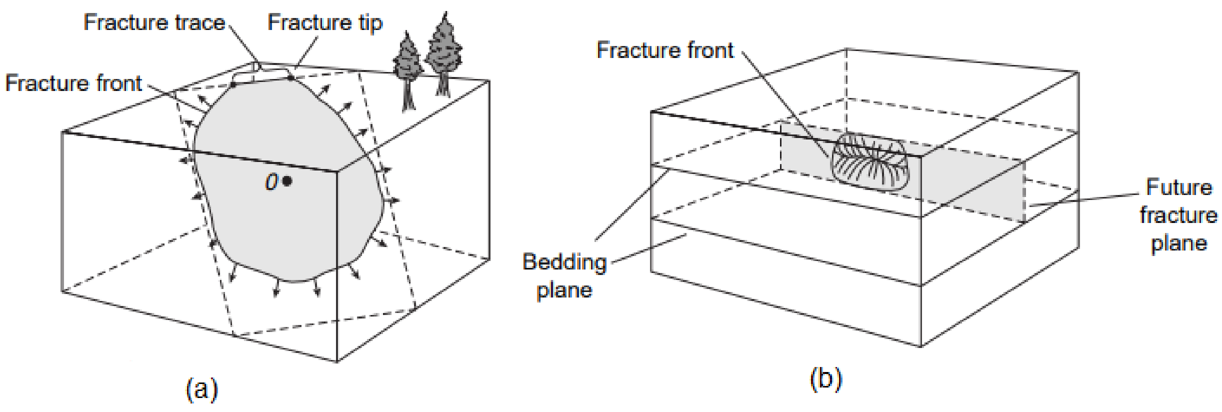

Figure 1 (a) A block diagram illustrating that a fracture surface terminates within the limits of a rock body. The top surface of the block is the ground surface; erosion exposes the fracture trace. Note that the trace of the fracture on the ground surface is a line of finite length (with a fracture tip at each end). The arrows indicate that this particular fracture grew radially outward from an origin, labeled O, the dot in the center of the fracture plane. (b) A blade fracture which has propagated in a sedimentary layer and terminates at the bedding plane.

What is brittle deformation?

To understand brittle deformation, we need to look at the atomic structure of materials. Solids are composed of atoms or ions that are connected to one another by chemical bonds that can be thought of as tiny springs. Each chemical bond has an equilibrium length, and the angle between any two chemical bonds connected to the same atom has an equilibrium value (Figure 2a). During elastic strain, the bonds holding the atoms together within the solid stretch, shorten, and/or bend, but they do not break (Figure 2b)! When the stress is removed, the bonds return to their equilibrium conditions and the elastic strain disappears. In other words, elastic strain is fully recoverable. Recall that this elastic property of solids explains the propagation of earthquake waves though Earth. Rocks cannot accumulate large elastic strains; you certainly cannot stretch a rock to twice its original length and expect it to spring back to its original shape. At most, a rock can develop a few percent strain by elastic distortion. If the stress applied to a rock is greater than the stress that the rock can accommodate elastically, then one of two changes can occur: the rock deforms in a ductile manner, or the rock deforms in a brittle manner. If the stress becomes large enough to stretch or bend chemical bonds so much that the atoms are too far apart to attract one another, then the bonds break, resulting in either formation of a fracture (Figure 2c) or slip on a preexisting fracture. In contrast to elastic strain, brittle deformation is nonrecoverable, meaning that the distortion remains when the stress is removed. Again, we will use earthquakes as an example; in this case, elastic strain is exceeded at the focus, resulting in failure and displacement. The pattern of breakage during brittle deformation depends on stress conditions and on material properties of the rock, so brittle deformation does not involve just one process. For purposes of our discussion, we divide brittle deformation processes into four categories that are Cataclastic flow, Frictional sliding, Shear rupture and Tensile cracking Figure 3.

Tensile cracking: Stress Concentration and Griffith Cracks

In Figure 4 we illustrate a crack in rock on the atomic scale. One way to create such a crack would be for all the chemical bonds across the crack surface to break at once. In this case, the tensile stress necessary for this to occur is equal to the strength of each chemical bond multiplied by all the bonds that had once crossed the area of the crack. If you know the strength of a single chemical bond, then you can calculate the stress necessary to break all the bonds simply by multiplying the bond strength by the number of bonds. Using realistic values for the elasticity (Young's modulus, E) and small strain results in a theoretical strength of rock that is thousands of megapascals. Measurement of rock strength in the Earth's crust shows that tensile cracking occurs at crack-normal tensile stresses of less than about 10 MPa, when the confining pressure is low, a value that is hundreds of times less than the theoretical strength of rock. Keeping the concept of theoretical strength in mind, we therefore face a paradox: How can natural rocks fracture at stresses that are so much lower than their theoretical strength?

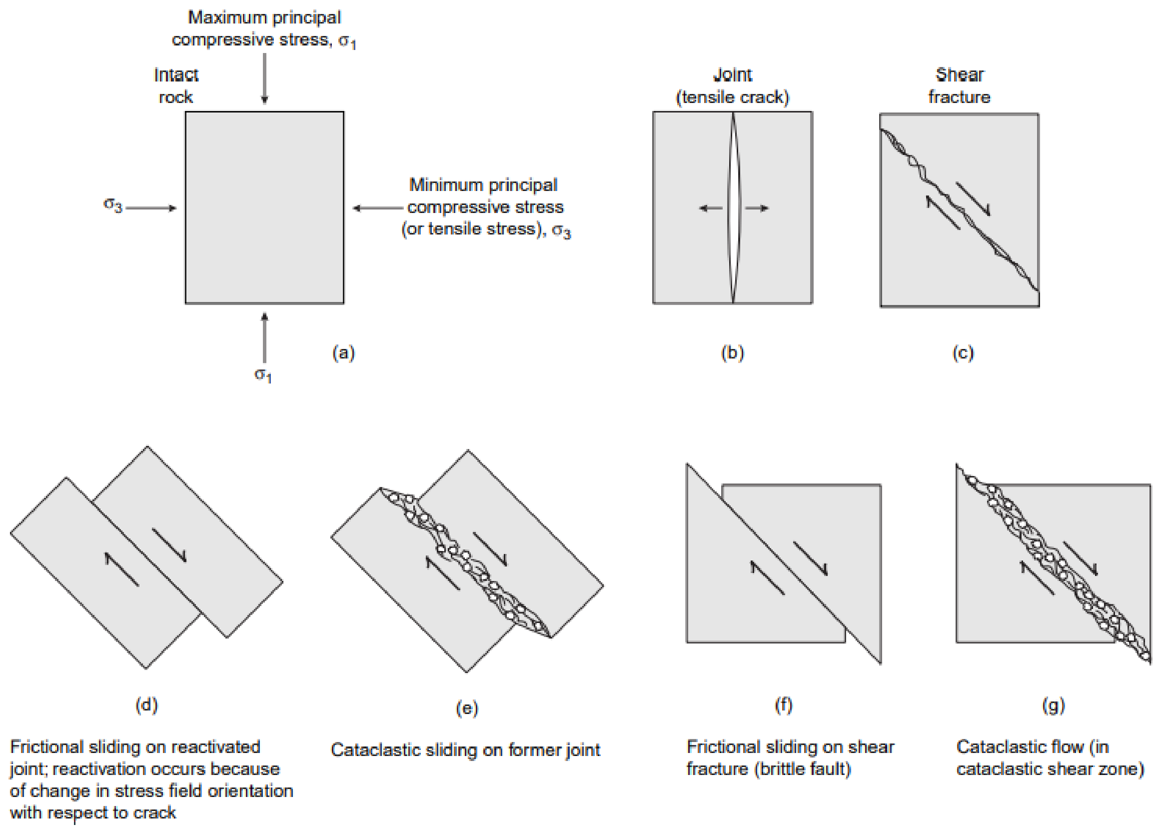

Figure 3 Types of brittle deformation (a) Orientation of the remote principal stress directions with respect to an intact rock body. (b) A tensile crack, forming parallel to σ1 and perpendicular to σ3 (which may be tensile). (c) A shear fracture, forming at an angle of about 30° to the σ1 direction. (d) A tensile crack that has been reoriented with respect to the remote stresses and becomes a fault by undergoing frictional sliding. (e) A tensile crack which has been reactivated as a cataclastic shear zone. (f) A shear fracture that has evolved into a fault. (g) A shear fracture that has evolved into a cataclastic shear zone.

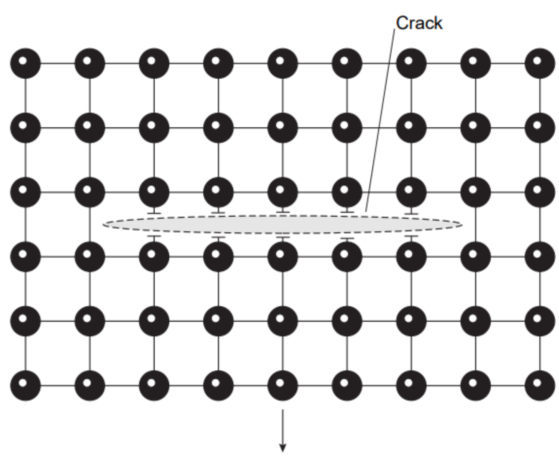

Figure 4 A cross-sectional sketch of a crystal lattice (balls are atoms and sticks are bonds) in which there is a crack. The crack is a plane of finite extent across which all atomic bonds are broken

The first step toward resolving the strength paradox came when engineers studying the theory of elasticity realized that the remote stress (stress due to a load applied at a distance from a region of interest) gets concentrated at the sides of flaws (e.g., holes) inside a material. For example, in the case of a circular hole in a vertical elastic sheet subjected to tensile stress at its ends (Figure 5a), the local stress (i.e., stress at the point of interest) tangent to the sides of the hole is three times the remote stress magnitude (σ_r). The magnitude of the local tangential stress at the top and bottom of the hole equals the magnitude of the remote stress, but is opposite in sign (i.e., it is compressive). If the hole has the shape of an ellipse instead of a circle (Figure 5b), the amount of stress concentration, C, is equal to 2b/a + 1, where a and b are the short and long axes of the ellipse, respectively. Thus, values for stress concentration at the ends of an elliptical hole depend on the axial ratio of the hole: the larger the axial ratio, the greater the stress concentration. For example, at the ends of an elliptical hole with an axial ratio of 8:1, stress is concentrated by a factor of 17, and in a 1 μm × 0.02 μm crack the stress is magnified by a factor of ∼100!

With this understanding in mind, A. W. Griffith, in the 1920s, took the next step toward resolving the strength paradox when he applied the concept of stress concentration at the ends of elliptical holes to fracture development. Griffith suggested that all materials contain preexisting microcracks or flaws at which stress concentrations naturally develop, and that because of the stress concentrations that develop at the tips of these cracks, they propagate and become larger cracks even when the host rock is subjected to relatively low remote stresses.

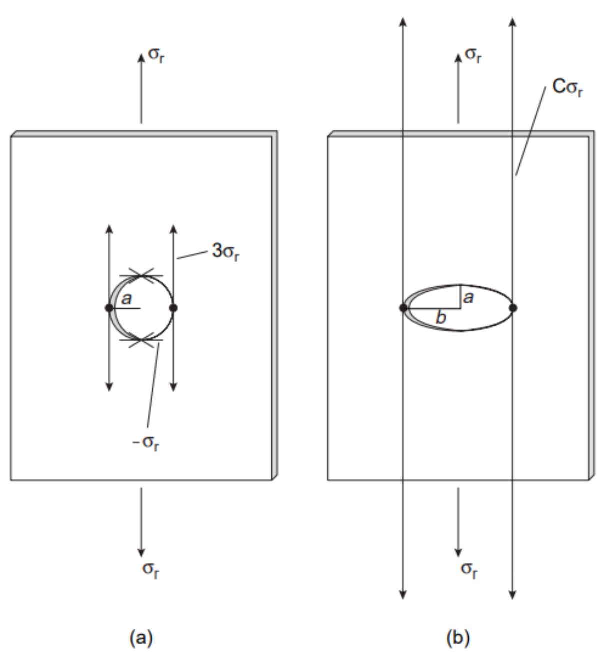

Figure 5 Stress concentration adjacent to a hole in an elastic sheet. If the sheet is subjected to a remote tensile stress at its ends (σr), then stress magnitudes at the sides of the holes are equal to Cσr, where C, the stress concentration factor, is (2b/a) + 1. (a) For a circular hole, C = 3. (b) For an elliptical hole, C > 3

He discovered that in a material with cracks of different axial ratios, the crack with the largest axial ratio will most likely propagate first. In other words, stress at the tips of preexisting cracks can become sufficiently large to rupture the chemical bonds holding the minerals together at the tip and cause the crack to grow, even if the remote stress is relatively small. Preexisting microcracks and flaws in a rock, which include grain-scale fractures, pores, and grain boundaries, are now called **Griffith cracks** in his honor. Thus, we resolve the strength paradox by learning that rocks in the crust are relatively weak because they contain Griffith cracks. Griffith's concept provided useful insight into the nature of cracking, but his theory did not adequately show how factors such as crack shape, crack length, and crack orientation affect the cracking process. In subsequent years, engineers developed a new approach to studying the problem. In this approach, called linear elastic fracture mechanics, we assume that cracks in a material have nearly infinite axial ratio (defined as long axis/short axis), meaning that all cracks are very sharp. Linear elastic fracture mechanics theory predicts that, if factors like shape and orientation are equal, a longer crack will propagate before a shorter crack. We can examine how preexisting cracks affect the magnitude of stress necessary for tensile cracking in a simple experiment. Take a sheet of paper (Figure 6) and pull at both ends.

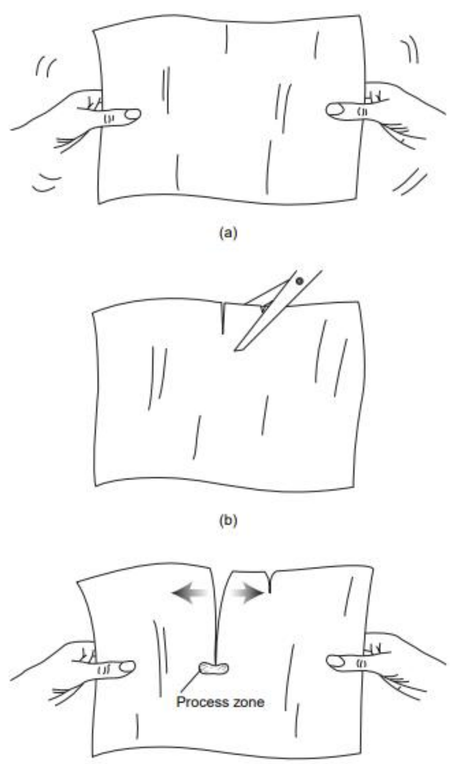

Figure 6 Illustration of a home experiment to observe the importance of preexisting cracks in creating stress concentrations. (a) An intact piece of paper is difficult to pull apart. (b) Two cuts, a large one and a small one, are made in the paper. (c) The larger preexisting cut propagates. In the shaded area, a region called the process zone, the plastic strength of the material is exceeded and deforms.

You have to pull quite hard in order for the paper to tear. Now make two cuts, one that is ∼0.5 cm long and one that is ∼2 cm long, in the edge of the sheet near its center, and pull again. The pull that you apply gets concentrated at the tip of the preexisting cuts, and at this tip the strength of the paper is exceeded. You will find that it takes much less force to tear the paper, and that it tears apart by growth of the longer preexisting cut. The reason that sharp cracks do not propagate under extremely small stresses is that the tips of real cracks are blunted by a crack-tip process zone, in which the material deforms plastically (Figure 6c). It is implicit in our description of crack propagation that the total area of a crack does not form instantaneously, but rather a crack initiates at a small flaw and then grows outward.

Exploring Tensile Crack Development

Let's consider what happens during a laboratory experiment in which we stretch a rock cylinder along its axis under a relatively low confining pressure (Figure 7a), a process called axial stretching. As soon as the remote tensile stress is applied, preexisting microcracks in the sample open slightly, and the remote stress is magnified to create larger local stresses at the crack tips. Eventually, the stress at the tip of a crack exceeds the strength of the rock and the crack begins to grow. If the remote tensile stress stays the same after the crack begins to propagate, then the crack continues to grow, and may eventually reach the sample's margins. When this happens, the sample fails, meaning it separates into two pieces that are no longer connected (Figure 7b).

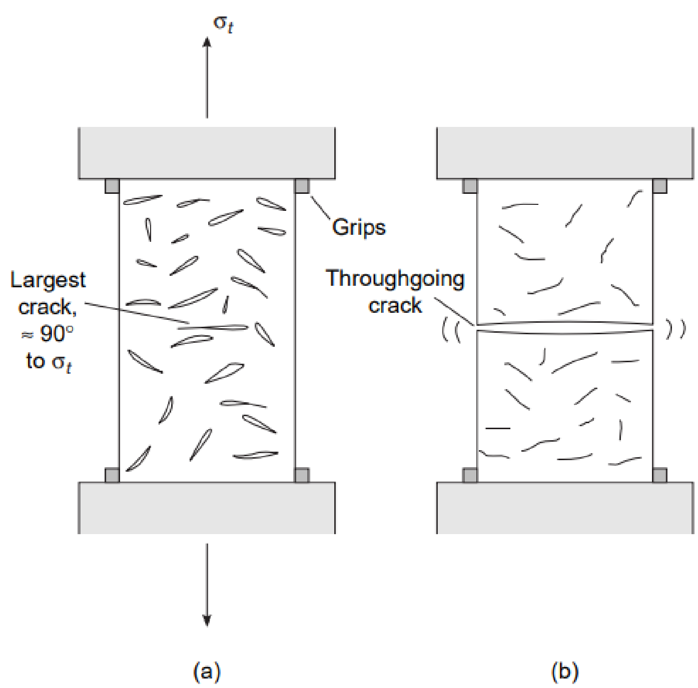

Figure 7 Development of a throughgoing crack in a block under tension. (a) When tensile stress (σ_t) is applied, Griffith cracks open up. (b) The largest, properly oriented cracks propagate to form a throughgoing crack.

We can also induce tensile fracturing by subjecting a rock cylinder to axial compression, under conditions of low confining pressure. Under such stress conditions mesoscopic tensile fractures develop parallel to the cylinder axis (Figure 8a), a process known as longitudinal splitting. Longitudinal splitting is similar to tensile cracking except that, in uniaxial compression, the cracks that are not parallel to the σ1 direction are closed, whereas cracks that are parallel to the compression direction can open up. To picture this, imagine an envelope standing on its edge. If you push down on the top edge of the envelope, the sides of the envelope pull apart, even if they were not subjected to a remote tensile stress (Figure 8b). In rocks, as the compressive stress increases, the tensile stress at the tips of cracks exceeds the strength of the rock, and the crack propagates parallel to the compressive stress direction. In the compressive stress environment illustrated in (Figure 8), the confining pressure required is very small; but tensile cracks can also be generated in a rock cylinder when the remote stress is compressive under higher confining pressure when adding fluid pressure in pores and cracks of the sample (i.e., the pore pressure; Figure 9). The uniform, outward push of a fluid in a microcrack can have the effect of creating a local tensile stress at crack tips, and thus can cause a crack to propagate. We call this process hydraulic fracturing. As soon as the crack begins to grow, the volume of the crack increases, so if no additional fluid enters the crack, the fluid pressure decreases. Crack propagation ceases when pore pressure drops below the value necessary to create a sufficiently large tensile stress at the crack tip, and does not begin again until the pore pressure builds up sufficiently. Therefore, tensile cracking driven by an increase in pore pressure typically occurs in pulses.

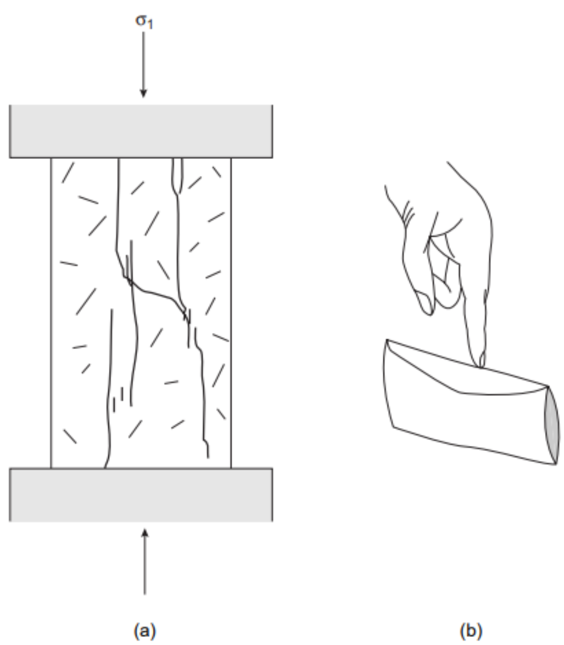

Figure 8 (a) A cross section showing a rock cylinder with mesoscopic cracks formed by the process of longitudinal splitting. (b) An "envelope" model of longitudinal splitting. If you push down on the top of an envelope (whose ends have been cut off), the sides of the envelope will move apart.

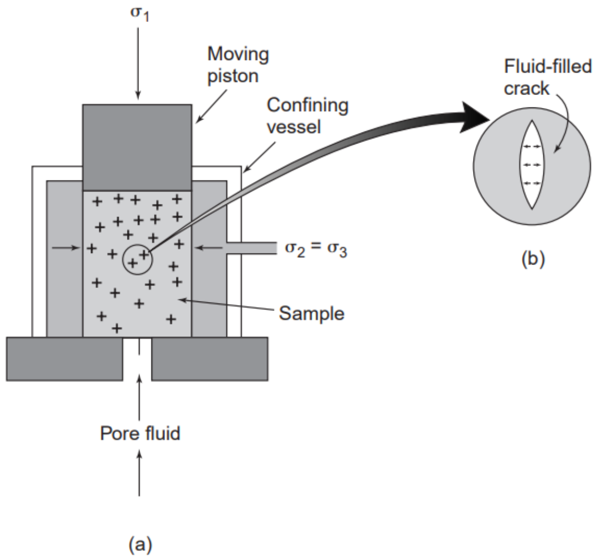

Figure 9 (a) Cross-sectional sketch illustrating a rock cylinder in a triaxial loading experiment. Fluid has access to the rock cylinder and fills the cracks. (b) A fluid-filled crack that is being pushed apart from within by pore-fluid pressure.

Modes of Crack-Surface Displacement

Before leaving the subject of Griffith cracks, we need to address one more critical issue, namely, the direction in which an individual crack grows when it is loaded. So far, we have limited ourselves to cracks that are perpendicular to a remote stress. But what about cracks in other orientations with respect to stress, and how do they propagate? Materials scientists identify three configurations of crack loading. These configurations result in three different modes of crack-surface displacement (Figure 10). Note that the “displacement” we are referring to when describing crack propagation is only the infinitesimal movement initiating propagation of the crack tip and is not measurable mesoscopic displacement as in faults. During Mode I displacement, a crack opens very slightly in the direction perpendicular to the crack surface, so Mode I cracks are tensile cracks. They form parallel to the principal plane of stress that is perpendicular to the $\sigma 3$ direction, and can grow in their plane without changing orientation. During Mode II displacement (the sliding mode), rock on one side of the crack surface moves very slightly in the direction parallel to the fracture surface and perpendicular to the fracture front. During Mode III displacement (the tearing mode), rock on one side of the crack slides very slightly parallel to the crack but in a direction parallel to the fracture front. Although shear-mode cracks appear similar to mesoscopic faults, Mode II (and Mode III) cracks are not simply microscopic equivalents of faults.

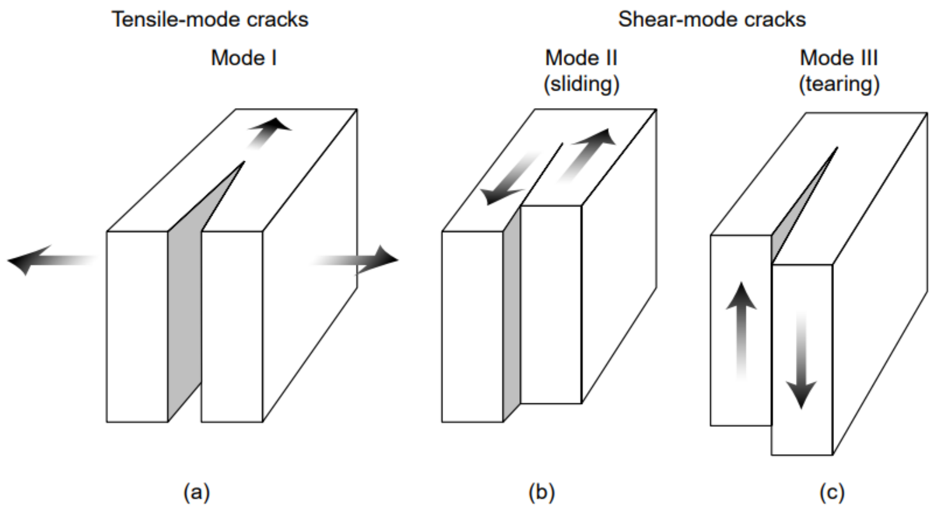

Figure 10 Block diagrams illustrating the three modes of crack surface displacement: (a) Mode I, (b) Mode II, (c) Mode III. Mode I is a tensile crack, and Mode II and Mode III are shear-mode cracks.

References

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