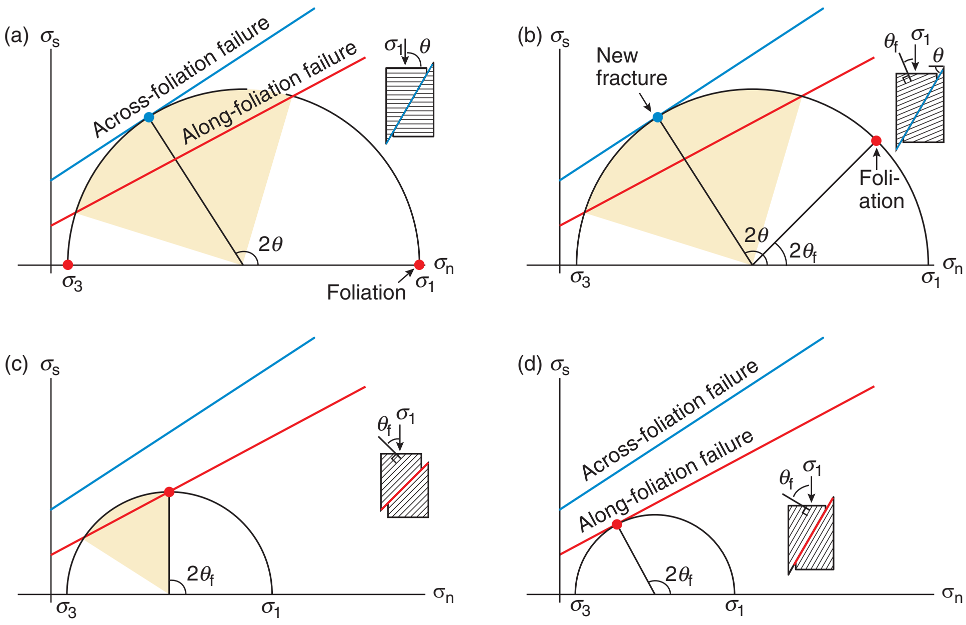

Figure 7.22 Illustration of the role of a preexisting foliation, for constant σ_3. (a) σ_1 acting perpendicular to the foliation, in which case differential stress builds up until the Mohr circle touches the upper envelope and across-foliation failure occurs. Colored sector indicates the range of orientations for along-foliation failure. (b) σ_1 at a high angle to the foliation, still too high for foliation-parallel failure (foliation still outside of the colored sector). (c) σ_1 at 45° to the foliation, causing foliation-parallel failure. Sector indicates the range of foliation orientations where along-foliation failure would occur for this particular state of stress. (d) The angle between σ_1 and the foliation that gives failure at the lowest possible differential stress. This is the weakest direction of a foliated rock.

Whether a rock fails along a weak preexisting fabric or fracture depends on the orientation of the fracture relative to the stress field.

The Mohr diagram and the failure envelopes discussed above only consider confining and differential stress and do not take into consideration σ_2. Experiments show that the influence of σ_2 is small and most pronounced when two of the stress axes are equal in size. For a vertical σ_1, the dip of the shear fracture is lowest when $\sigma_2 = \sigma_1$ and highest when $\sigma_2 = \sigma_3$. For foliated rocks where the foliation does not contain the intermediate principal stress axes, the influence of σ_2 is greater. In this case the resolved normal and shear stress on the foliation depends on all three principal stresses.

Temperature has a major influence on rheology in the plastic regime, but its influence within the brittle regime is relatively small for most common minerals. It does however control the range of the brittle regime as increasing temperature lowers the von Mises yield stress (lowers the yield point or the stress at which rock flows plastically).

An interesting laboratory observation is related to sample size: as the size of the sample increases, its strength is reduced. The reason for this somewhat surprising finding is simply that large samples contain more microfractures than small samples. Because microfractures differ in length and shape, a large sample is likely to contain some microfractures that have a shape that causes larger stress concentrations than any of those in a smaller sample of the same rock.

During a rock experiment, a large sample is likely to fracture before a smaller one.

The dependence on scale is even more pronounced at larger scales. Think of all the joints, faults and other weak structures in the crust that will be activated before the strength of the rock itself is reached. Such weak structures control the strength of the brittle crust, which means that the upper crust is not by far as strong as suggested by experimental testing of unfractured samples in the laboratory. This brings us to another important topic; the reactivation of brittle fractures by frictional sliding.

Growth and morphology of fractures

Shear fractures cannot propagate in their own plane, but rather spawn new tensile cracks (wing cracks) according

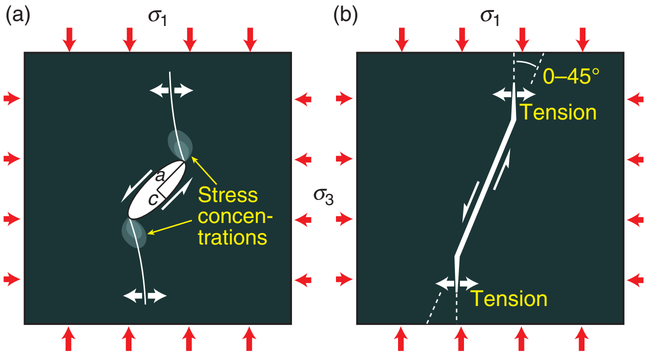

Figure 7.23 (a) Griffith crack modeled as an elliptical void. Tensile stress concentrates near the crack tips (compare with Figure 7. 20). (b) A critically stressed Griffith crack at the tip of a shear fracture. The Griffith crack is oriented between 0 and 45° with respect to σ_1, depending on the ratio σ_1/σ_3. Note that tensile stress develops near the crack tips in spite of the overall compressional stress, that crack growth is accommodated by sliding along the main crack, and that the crack grows toward parallelism with σ_1.

to Griffith's theory (Figure 7.23 and Box 7.3) or develop by activation of already existing extension fractures. In contrast, extension fractures may propagate into long structures. Ideally, an extension fracture will grow radially from a nucleation point so that at any point the propagation front (tipline) has the shape of an ellipse (Figure 7.24a, b). The rate of propagation increases after initiation, and the joint surface gets rougher until it propagates so fast that the stress readjustments or stress oscillations at the crack tip cause it to bend. In detail, the tip bifurcates and off-plane microcracks form because of high stress and/or local heterogeneities in the tip zone. The result is long, narrow planes slightly oblique to the main fracture surface named **hackles** (Figure 7.24c), and the hackles form **plumose** (featherlike) **structures**. Plumose structures reflect the propagation direction along the plume axis, as shown in Figure 7.25.

Locally the main fracture may enter an area with a different stress orientation. This would typically be a bedding interface or some other boundary between two rock types of different mechanical properties, in which case a series of twisted joints or **twist hackles** form in what is called a **fringe zone**. Twist hackles tend to be oriented **en echelon** because of the shear component on the main fracture locally imposed by the new orientation of σ_3. The twist hackles try to orient perpendicular to σ_3, hence the twisting (Figure 7.26).

Extension fractures tend to grow in pulses. Each propagation pulse tends to end with an out-of-plane propagation with a slowing down or complete arrest until

BOX 7.3 FRACTURE GROWTH AND WING CRACKS

One of the peculiarities of rock mechanics is the fact that, even though a deforming sample develops through-going shear fracture(s) that make an acute (~30°) angle to σ_1, shear fractures cannot grow in their own plane. Instead, a Mode I fracture forms parallel with σ_1. Such fractures are known as wing cracks or edge cracks. In three dimensions wing cracks (Mode I) will form along both the Mode II and Mode III edges of the main fracture.

This development is in agreement with theoretical stress considerations. But how does the shear fracture propagate from this stage? The general answer is that the Mode I wing cracks are broken by a new shear fracture – a process that keeps repeating as the main fracture grows. The result is a zone of minor fractures along and around the main shear fracture, a sort of damage zone akin to that defined for faults in Chapter 8.

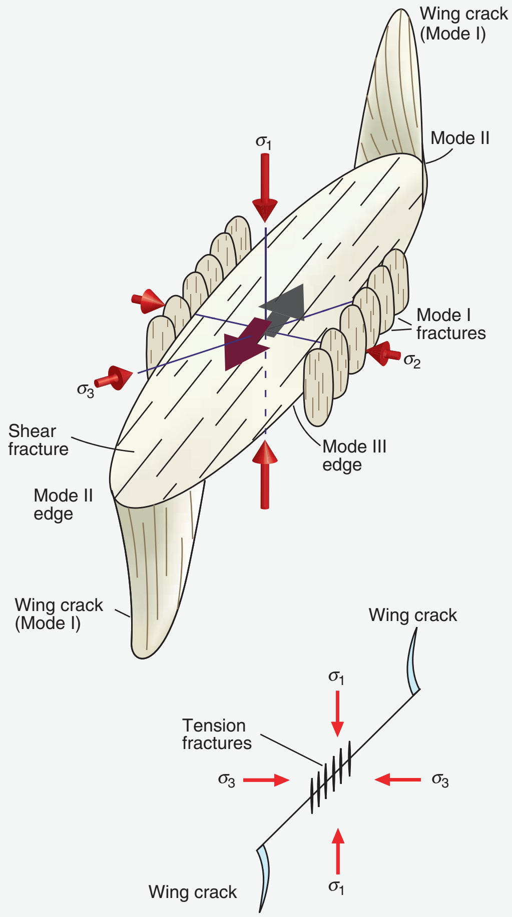

Illustration based on Scholtz (1990).

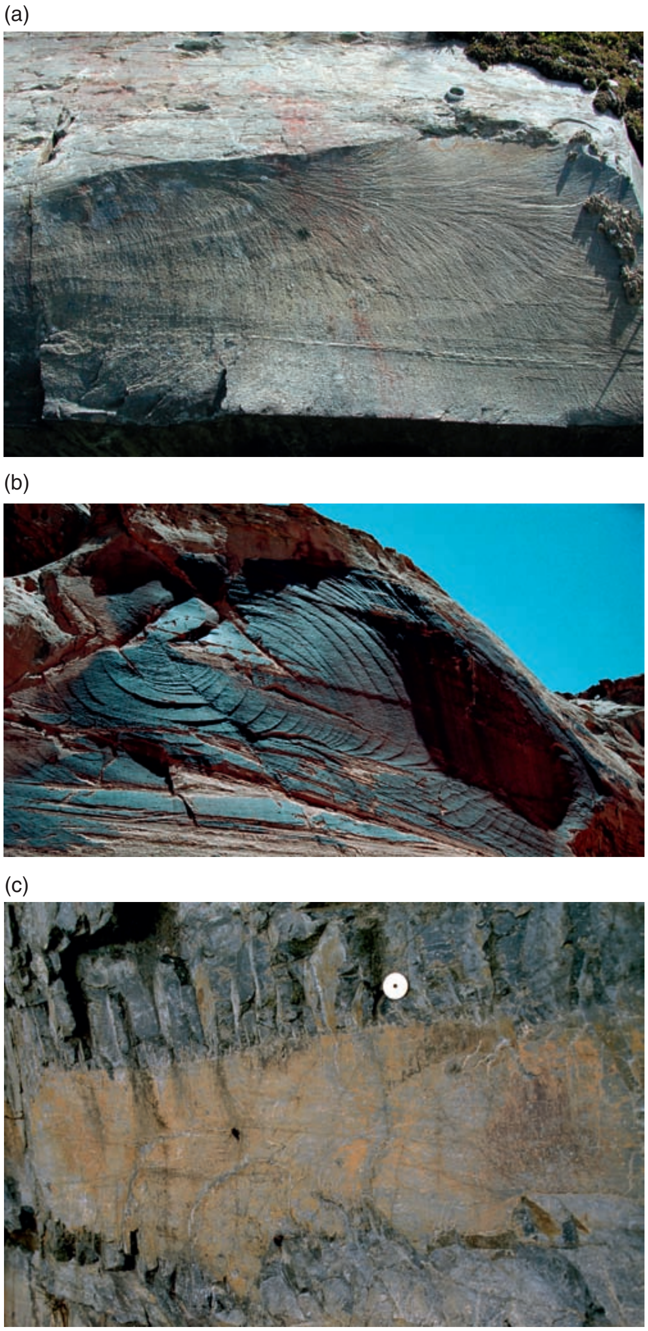

Figure 7.24 (a) Arrest lines and plumose structures in meta-greywackes from Telemark, Norway. Note the faint arrest lines oriented perpendicular to the plumose hackles. (b) Elliptically arranged arrest lines in the Navajo Sandstone, Utah. This sandstone is too coarse-grained for the plumose pattern to show up. (c) En-echelon hackle fringes (twist hackles) along a fracture in meta-rhyolite in the Caledonides of West Norway.

enough energy has built up to initiate the next pulse. Ribs are thus locations of minimum propagation velocity and form parabolic (elliptic in massive rocks) irregularities sometimes referred to as **arrest lines**. Ribs are perpendicular to the plumose hackles (Figures 7.24a and 7.25) and

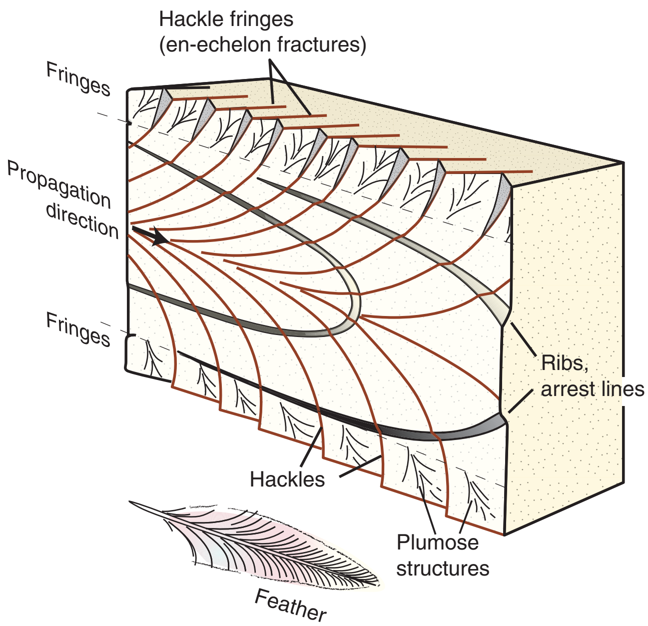

Figure 7.25 Schematic illustration of structures characteristic of joint surfaces. Based on Hodgson (1961).

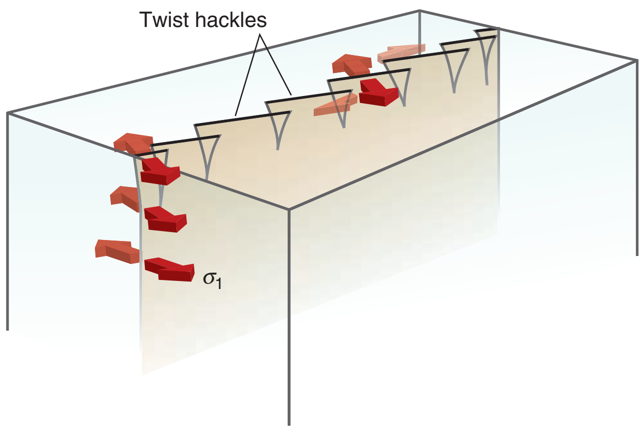

Figure 7.26 The twisting of extension fractures as they reach an interface with a mechanically different rock layer. Note the parallel twisting of σ_1 and the fractures (hackles). Compare with the hackles illustrated in Figure 7.24c.

together these structures provide unique information about the growth history of extension fractures. Plumose structures are characteristic for joints in fine-grained rocks such as siltstones, while arrest lines are also commonly seen in coarser-grained lithologies such as sandstones and granites.

7.5 Fracture termination and interaction

Studies of shear fracture terminations reveal that they sometimes split into two or more fractures with new orientations and, as indicated in Figure 7.27, a spectrum of different fracture geometries can been found. We have