JOINTS

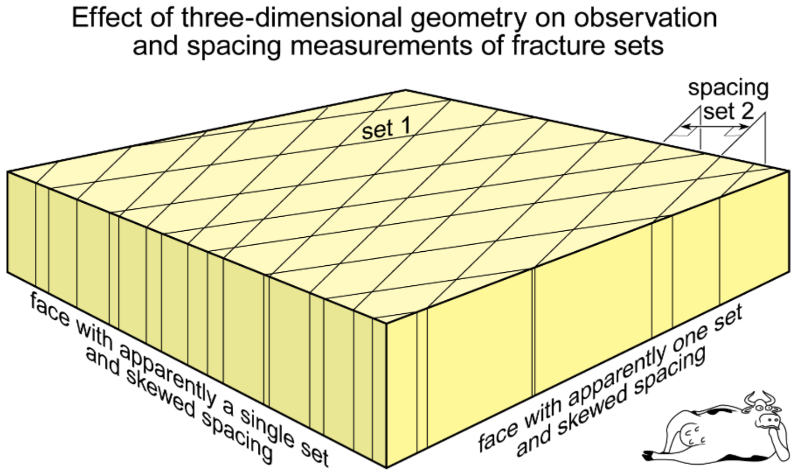
Joints (also termed **extensional fractures**) are planes of separation on which no or undetectable shear displacement has taken place. The two **walls** of the resulting tiny opening typically remain in tight (matching) contact. Joints may result from regional tectonics (i.e. the compressive stresses in front of a mountain belt), folding (due to curvature of bedding), faulting, or internal stress release during uplift or cooling. They often form under high fluid pressure (i.e. low effective stress), perpendicular to the smallest principal stress.
The **aperture** of a joint is the space between its two walls measured perpendicularly to the mean plane. Apertures can be open (resulting in permeability enhancement) or occluded by mineral cement (resulting in permeability reduction). A joint with a large aperture (> few mm) is a **fissure**.
The **mechanical layer** thickness of the deforming rock controls joint growth. If present in sufficient number, open joints may provide adequate porosity and permeability such that an otherwise impermeable rock may become a productive **fractured reservoir**. In quarrying, the largest block size depends on joint frequency; abundant fractures are desirable for quarrying crushed rock and gravel.

Joint sets and systems

Joints are ubiquitous features of rock exposures and often form families of straight to curviplanar fractures typically perpendicular to the layer boundaries in sedimentary rocks. A **set** is a group of joints with similar orientation and morphology. Several sets usually occur at the same place with no apparent interaction, giving exposures a blocky or fragmented appearance. Two or more sets of joints present together in an exposure compose a **joint system**. Joint sets in systems commonly intersect at constant **dihedral** angles. They are conjugate for dihedral angles from 30 to 60°, orthogonal when the dihedral angle is nearly 90°

Geometry

The geometry of joint systems refers to the orientation (plotted on stereonets and rose-diagrams), the scale, the shapes and trajectories, the spacing, the aperture, the intersections and terminations of the studied joints. The mean orientation and orientation distribution, spacing and relative chronology are general characters used to define joint sets. In this respect, a three-dimensional observation is essential to avoid skewed sampling measurements due to simple geometrical reasons.



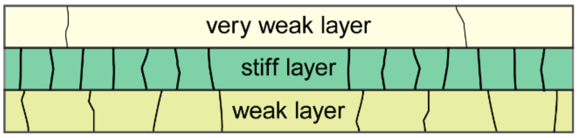
Bedding-contained joints terminate at the top and bottom of beds.
Systematic joints are characterized by a roughly planar geometry; they have relatively long traces and typically form sets of approximately parallel and almost equally spaced joints.
Non-systematic joints are usually short, curved and irregularly spaced. They generally terminate against systematic joints.

Spacing

The sizes and **spacing** (the average orthogonal distance between neighboring fracture planes) are essential characteristics of joint sets. In isotropic rocks (e.g. granite) joint spacing follows an approximately log-normal **frequency** (the number of joints occurring within a unit length) distribution. In anisotropic (layered) rocks, joint spacing differs according to several parameters.

Bed thickness

For the same lithology, joints are more closely spaced in thinner beds. This is because the formation of joints relieves tensile stress in the layer over a lateral distance proportional to the joint length. Since joints end at layer boundaries, which are rock discontinuities, the longer joints in thicker layers need to be spaced less frequently.



Spacing of bedding-parallel joints
according to competence (elastic modulus) of layers

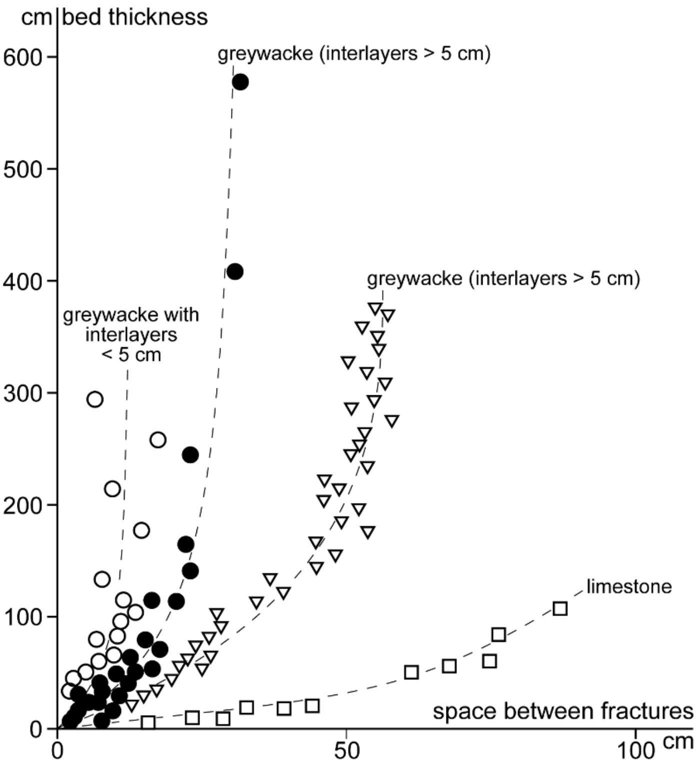
Much work has documented linear relationships between average joint spacing, D, and bed thickness, T:

$$D = \alpha T$$

The slopes α are a function of lithology and, by inference, of mechanical properties. However, this linear relationship might be valid for beds less than 1.5m thick. The slope may change for thicker beds. A continuous curve with a positive slope and a negative second derivative:

$$d^2D/dT^2 < 0$$

acceptably fits all data.



Relationship between fracture separation and bed thickness
for different rock types from different places
from Ladeira & Price 1981 *Journal of Structural Geology* **3(2)** 179-183

However, systematic investigations have shown that the thickness of incompetent interlayers influences fracture separation within competent layers. Spacing is wider where interlayers are thicker than a critical value assessed to be ca 5 cm; conversely, fractures are closer to each other where weak interlayers are thinner than 5 cm.
Spacing scaled with layer thickness is a tool to map lithological contacts, particularly in air-photo interpretation or in the surface mapping of heavily weathered or inaccessible exposures. Spacing may also reveal differences in the joint systems at limb and hinge positions on large folds or different distances from large faults.

Lithology

Stronger, more brittle rocks have more closely spaced joints than weaker rocks. Similarly, rocks with low tensile strength show more joints than stiffer lithologies, because the strain is the same along layers of different types. Yet, higher stresses are required to achieve the same amount of strain in the stronger layers. Therefore, strong layers fracture more frequently. However, this response is particularly sensitive to local pore fluid pressure.

Structural position and strain

The structural position (particularly within folds) and the magnitude of extensional strain also control joint spacing.

Why joints are evenly spaced?

The regular spacing of joints has met several explanations, none of which having been established as a proven mechanism. Yet, nearly all infer that joints form in sequence. Among these hypotheses:

Pore pressure / porosity interaction

When a joint forms, fluids flow into the fracture and the pore pressure in the adjoining rock diminishes. The local Mohr circle moves away from the failure envelope, and no fracture is possible near the initial one. Another fracture can only form beyond the volume of rock with reduced pore pressure. The minimum spacing thus depends on the permeability of the rock. This minimum distance would be the measured finite spacing of joint sets.

Exercise

Explain with a Mohr diagram and any given permeability level how pore pressure variations may control joint spacing.

Stress shadow

When a joint forms, it relieves the tensile stress on either side of the fracture plane that becomes a zero-stress surface. Stress builds up gradually away from the fracture until it reaches the remote stress level. The next fracture can only form beyond the volume of rock with reduced tensile stress (the stress shadow). The minimum spacing thus depends on the width of stress shadows, hence on elastic properties of the rock. The uniform spacing is ruled by adjacent stress shadows with joints in the middle. Stress shadows are larger for longer joints.

Inter-layer forces

Each layer is submitted to forces transmitted by adjacent layers. The differential strain between layers exerts tensile stresses in the more competent ones (a process invoked for boudinage). Spacing between joints is determined by the length of the layer necessary to build up stresses to the tensile strength level of the concerned lithology.

Joint patterns

There are five main arrangements:
- Parallel sets are curved or straight
- Fans sets along fold or intrusion crests
- Radiate sets around intrusion centers

- Concentric sets around intrusion and collapse centers (cone, ring or cylindrical)
- Polygonal sets as columnar or prismatic.

Master joints

Joints that have dimensions ranging from tens of centimeters to hundreds of meters and repeat distances of several centimeters to tens of meters are **master joints**. Besides, most rocks contain numerous inconspicuous joints of smaller size and closer spacing, some of them, the **microjoints** or **microfractures,** visible only in thin section under the microscope.

Joint orientation

The permeability of a fractured reservoir is often highly anisotropic because joints generally form in sub-parallel sets. However, more than one fracture set may be present, resulting in a complex **fracture network**. Knowledge of the spatial distribution and orientation of joints is therefore required to optimize the development of fractured reservoirs.

Joints are measured on a sampling station. The technique aims at measuring their orientation and the fracture density.

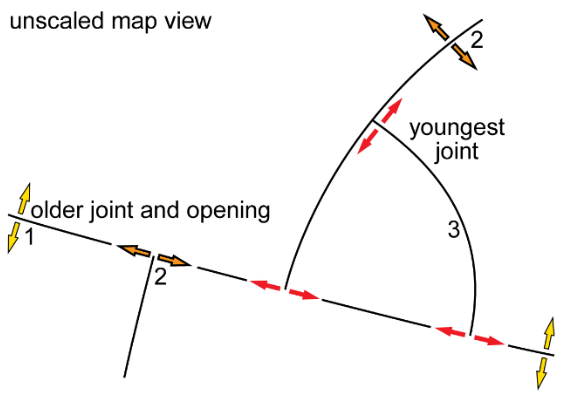
- A circle is drawn with a piece of chalk attached to a string of determined length (r = radius) on a perfectly exposed surface.
- One measures the orientation (strike and dip) and length L of each fracture within the circle, marking with the chalk each measured fracture to avoid data duplication. Where dips are not obtained, such as on photographs, orientation data are presented on a rose diagram: Joints in a given orientation sector (e.g. within 10°) are counted. A radial line is drawn in the median direction of each sector. The length of the line indicates the number of joints occurring in the corresponding sector.
- The joint density is the cumulative fracture length over the circle's area:

$$\rho_f = \sum L / \pi.r^2$$

In three dimensions, this density (intensity) is the surface area of joint per unit volume of rock.

Relative timing of joint formation

Joints of the same generation have likely the same orientation. However, rocks experience different stress regimes during their history with the result that several fracture sets are superimposed on each other to produce a fracture network. Crosscutting relationships of different joint sets allow determining their relative age. Some rules help in this task.



Butting relationships between joint generations with stress perturbation close to pre-existing joints

- Early joints tend to be long and relatively continuous.
- Early joints arrest the propagation and modify the orientation of later ones.

- As more joint sets develop in the rock, modification of the stress orientation by the pre-existing fractures may result in a poor correlation between the orientation of late joints and the regional stress field responsible for their formation.

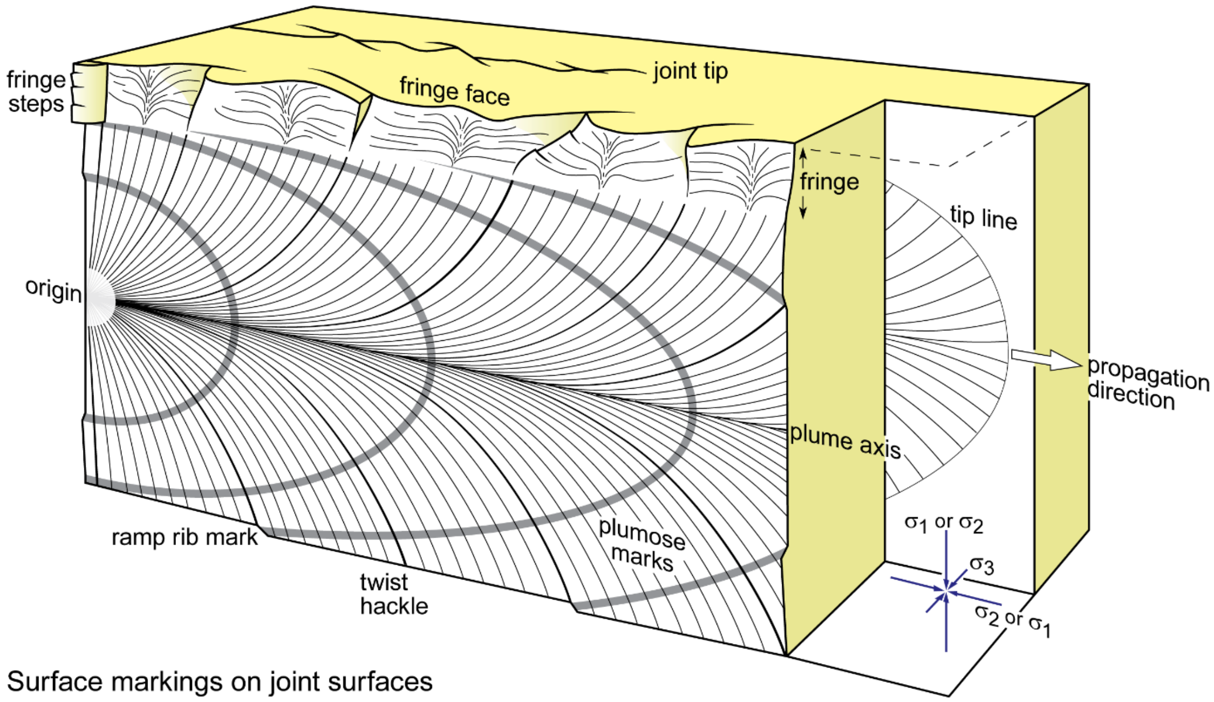
Younger joints must terminate against older joints because an extension fracture cannot propagate across another, older extension fracture. The principal stresses are reoriented near an early joint, which is a free surface unable to support shear stress within the rock and, therefore, is a principal plane of the stress ellipsoid. Consequently, later joints curve into orientation at a right angle to the earlier ones as they approach them and abut against these (**abutting relation**); younger joints are consequently shorter. This geometrical observation raises a point of caution: the orientation of secondary fractures does not directly reflect the regional stress field.

Joint anatomy

Joints normally are barren cracks or empty fissures but some may contain coatings. Narrow veins with infilling minerals, commonly quartz or calcite, are also extension fractures treated as joints.

Joint surfaces

Barren joints are characterized by clean, granular and jagged breaks. They are **conchoidal structures**, meaning that they are uneven surfaces with low relief convexities and concavities (like those of a clamshell) that do not follow any natural plane of separation. Such structures are typically seen when an amorphous material (glass, flint, obsidian, etc.) is fractured. Likewise, some joint surfaces display delicate ornaments falling into two groups: **plumose-marks**, the most common type, and **rib-marks**. Preservation of these delicate features specifies that the joint is not a shear fracture.



Surface markings on joint surfaces

Plumose structures

Plumose structures are aggregates of gentle curvilinear undulations (the **hackle marks**) that radiate from the point where the joint originated and fan outward from a generally straight, more rarely curved axial line, then resembling the shape and imprint of a feather. The **origin** commonly is some rock heterogeneity such as ripples on bedding planes or inclusions (concretion, nodule, clast, fossil, etc.) in beds. Hackles are often very fine near the joint origin, while the differential relief may amplify lengthwise towards the joint margin (the **fringe**). Hackles diverge sharply at angles of about 30° from

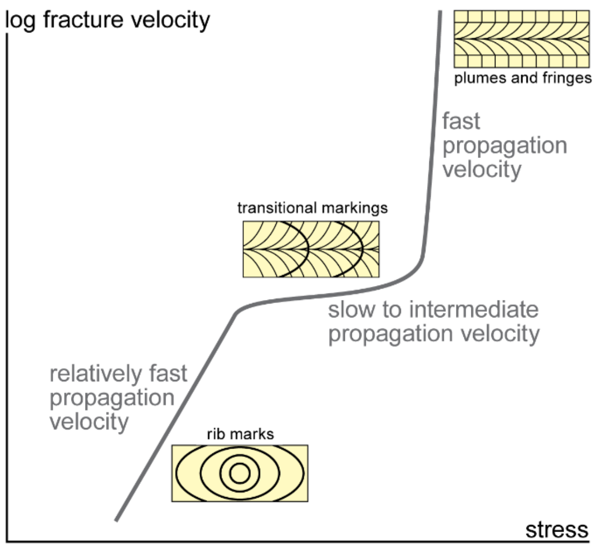
the central axis, gradually curving to angles of about 70° near the margins of the joint surface. The scale of plumose patterns seems to depend on the grain size of the rock.
Markings similar to plumose structures occur on fracture surfaces in glass and other brittle materials. They are interpreted as surface irregularities due to local variations in the propagation of the fracture front in terms of velocity and heterogeneities in the rock. Experiments show that the diverging rays of the plumose structures always remain parallel to the direction of propagation of the fracture. Thus, constructing lines at right angles to these rays yields the position and shape of the fracture front at different times of its evolution. The fracture fronts form a series of concentric ellipses, the center of which marks the site of fracture initiation.

Rib-marks

Rib-marks form a series of regular, concentric and arcuate changes or ramps in the orientation of the joint surface, giving cuspate, waveforms or rounded ridges or furrows. The central zone of rib-marks (the **mirror**) is often circular or elliptical. **Wallner lines** are similar to ribs but they occur as one or two sets oblique to the hackles.
Rib-marks represent changes in fracturing direction as the stress field changes. Experiments have shown that those that are sinusoidal in profile and smooth on their crests record the location at which the velocity of propagation of relatively fast-propagating fractures cutting through a solid material diminishes (the stress field is vibrating). Strongly asymmetric rib marks, with sharp crests and occasionally deviating from parallelism (**arrest marks**) are associated with slow crack propagation. They are old joint terminations, mapping successive temporary arrests of the crack front during repeated crack growth under recurring loading/unloading conditions.

Interpretation

Plumose and rib marks can be superposed on a joint plane and generally are orthogonal to each other. Such delicate features interlock on opposed faces of joints, and this precludes shear movement (hence, joints are mode 1 fractures parallel to the $(\sigma_1;\sigma_2)$ plane). Plumose and rib marks are a direct expression of the joint path because the edges of the fracture constantly twist and tilt as they advance. Plume axes develop parallel to the main propagation direction, commonly parallel to bedding. Considerations in linear fracture mechanics suggest that fracture velocity and /or stress intensity control these surface structures. The average propagation velocity has been measured to be half speed of sound. Experiments further suggest that propagation velocities of cracks with plumose ornamentation may exceed half the speed of sound.



Qualitative distribution of joint surface features
with respect to stress intensity and fracture propagation velocity
(adapted from Weinberger & Bahat, *Terra Nova* 2008, **20(1)**, 68-73)

Joint content - Secondary deposits

The secondary deposits along joints are **dilational** if the vein material occupies space between the two original fracture surfaces, or **non-dilational** if the vein material occupies space made available by **replacement** of the original rock outside the two original fracture surfaces.
In engineering practice, the mineralogy and texture of veins are important because joints with different fillings can have different mechanical properties governing the storage and flow of fluids. Then the fracture retains a certain degree of cohesive or tensile strength. In particular, if a fracture propagates slowly enough, healing may keep pace with propagation.

Joint termination and edges

At their **fringes**, joint surfaces split and twist into small out-of-plane fractures oblique (typically at 5-25°) and *en échelon* to the main joint. These **fringe faces** may individually bear secondary plumose patterns nearly orthogonal to the direction of propagation of the parent joint. They connect by curved or angular links. Small **fringe steps** parallel to the joint axis can segment fringe faces. Joint terminate with or without a change in orientation. Branching and abutting against another joint are other common descriptions.

Mechanics of jointing

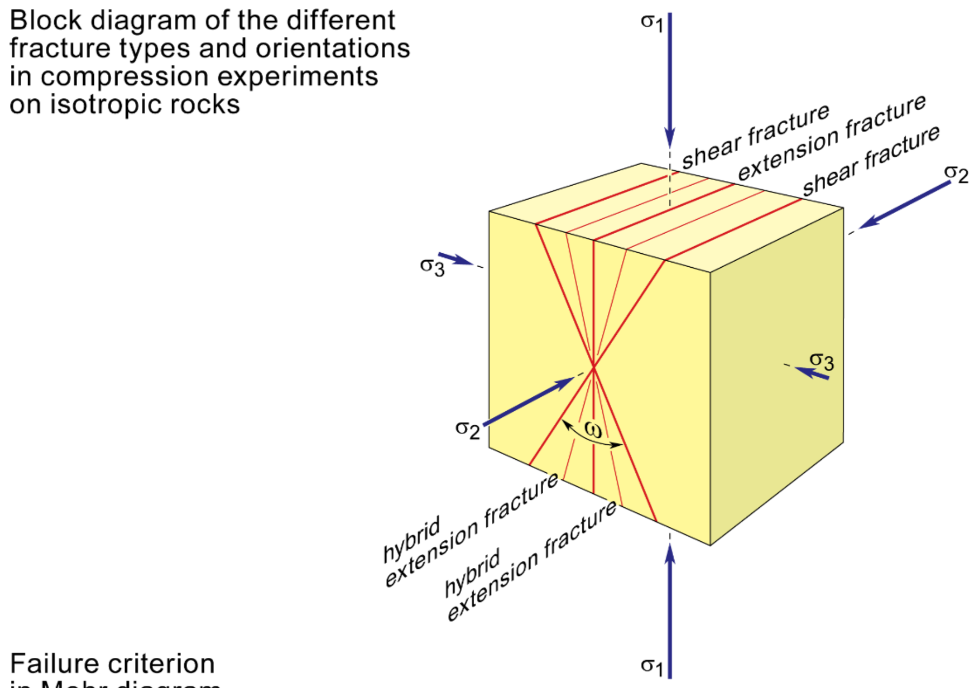
A genetic classification of joints is based on the size of inferred, imperceptible displacement related to the three principal stress axes of a region. If the total displacement is normal to the fracture surface, it is an **extension** or **dilatant joint** (mode 1 fracture). If the shear component has some finite, yet negligible value, the fracture called a **shear joint** is really a fault (modes 2 and 3 fracture), keeping in mind that the shear component may have accumulated after the formation of a former dilatant joint.

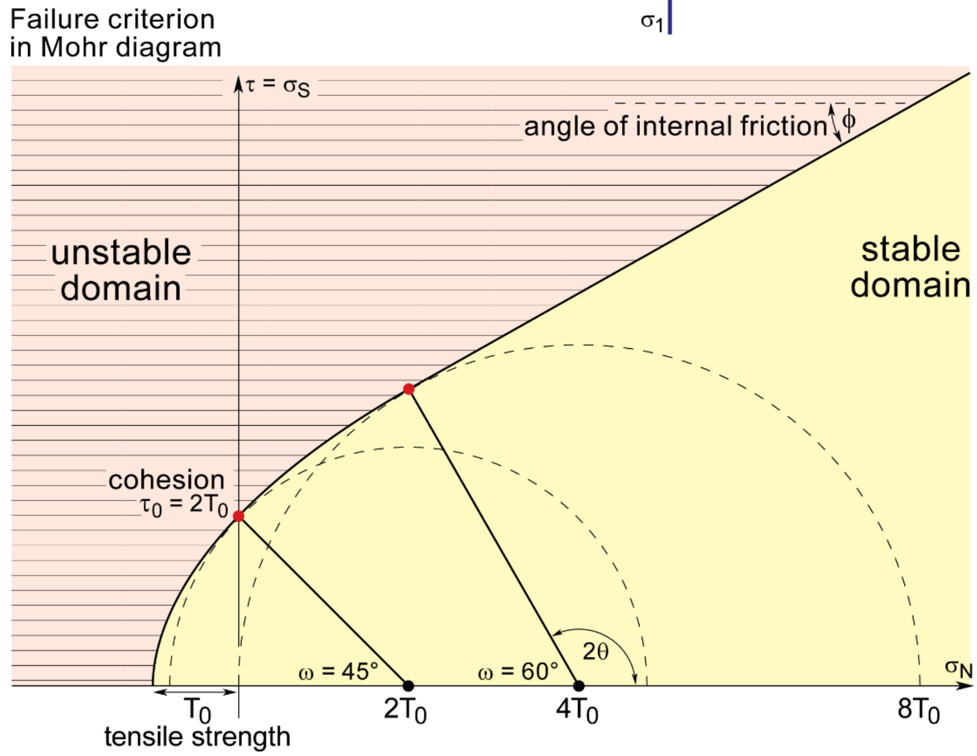
Relationship between joints and principal stresses

If we accept an average internal friction angle of 30° for the general Coulomb-type failure envelope, a 2θ angle of 135° on the Mohr diagram (i.e. the dihedral angle $\omega = 45°$ between conjugate fractures) divides fractures into subclasses. Conjugate fractures enclosing an angle of less than 45° involve negative σ_3 and σ_N; when $45 \leq \omega \leq 60°$, σ_N is positive although σ_3 remains negative. These relationships have led to a simple classification:

Failure mode	Class	$(\sigma_1 - \sigma_3)$	Dihedral angle
Tensile failure	Extension fracture	$< 4T_0$	0°
Hybrid shear failure	Extensional shear fracture	$4T_0 - 8T_0$	Up to 60°
Shear failure	Compressional shear fracture	$> 8T_0$	>60°

Note that there is a direct relationship between the magnitude of the differential stress $(\sigma_1 - \sigma_3)$ and the tensile strength T_0 of the rock, which is readily seen on the Mohr construction.

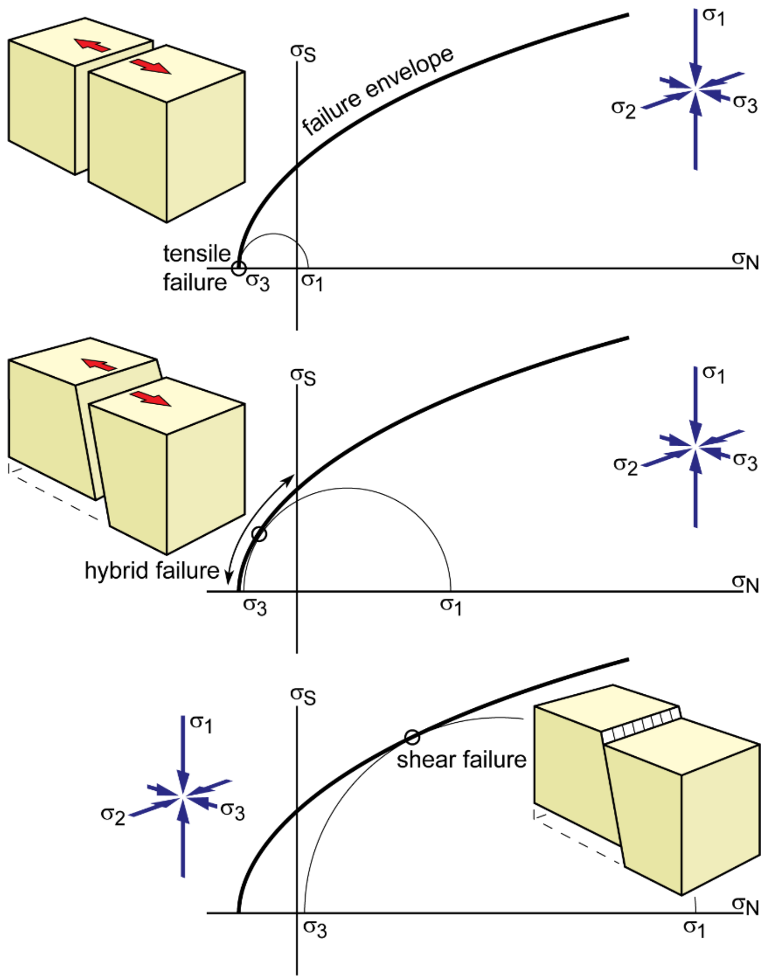




Extension joints

Linear elastic fracture mechanics predicts that the orientation of dilatant joints (genuine mode 1 fractures) in a relatively isotropic rock is controlled by the remote stress field at the time of fracture propagation: joints are gaping planes parallel to the maximum compressive stress σ_1 and perpendicular to the direction of the least principal stress σ_3. In other words, they form in the plane containing σ_1 and σ_2. Otherwise, there would be a shearing stress and a corresponding finite shear displacement on the joint plane. Triaxial experiments on brittle isotropic rocks confirmed this theoretical consideration. Thus, regionally consistent joint sets are taken as effective proxies for stress trajectories during joint growth: relatively closely spaced, parallel and linear joints suggest that the regional principal stress trajectories are rectilinear and remained parallel across the fractured area; alternatively, complex joint orientations are related to stress trajectory variability.

The pattern of dilatant joints is commonly T-shaped, the younger joint abutting the older one. Given suitable anisotropy of the tensile strength, it is, however, possible to get joints normal to σ_2 or even σ_1.



Hybrid joints

Hybrid joints show components of both extension and shear components. They are interpreted as failure surfaces initiated in the transition from tensile to shear failure. They form when the stress circle touches the (Griffith) failure envelope in the tensile (negative normal stresses) side of the Mohr diagram. Dihedral angles between conjugate hybrid fractures are typically smaller than between shear fractures (faults).

Shear joints

This term is unfortunate and ambiguous because shear joints actually are small faults. Conjugate "shear joints" generally define X, Y or V shapes. The acute bisector of these shapes is parallel to σ_1, unless these patterns represent unrelated crosscutting or abutting fractures.

Joint initiation

Geological observation indicates that joints initiate at rock "flaws" such as fossils, clasts, porphyroclasts, cavities, etc. Hard and soft inclusions perturb the remote stress field, hence are sites of joint nucleation in three ways:

- Amplification of a small remote tension so that the magnitude of local tensile stress at the flaw exceeds the tensile strength of the rock;
- Conversion of remote compression into local tension.
- Local tension due to pore pressure.

Amplification of remote tension

The reason is that flaws have different elastic properties than the rock. For such conditions, fracture mechanics consider a circular inclusion and a constant k, the ratio elastic shear modulus of the rock / elastic shear modulus of the inclusion. The rock and the inclusion have the same Poisson's ratio. A remote tension σ_{3r} induces a uniform tension σ_{3i} within the inclusion:

$$\sigma_{3i} = \sigma_{3r}\left[3k/(2k+1)\right]$$

and a tangential component σ_t at two opposite points of the inclusion boundary:

$$\sigma_t = \sigma_{3r}\left[3/(2k+1)\right]$$

If $k \to 0$ then the tangential stress is amplified by 3. Hard inclusions (high k ratio) amplify a remote tensile stress by factors up to 1.5 inside the inclusion while the tangential stress outside the inclusion is diminished. Near softer inclusions, the tangential stress is amplified, and for an open cavity or pore, this amplification is a factor of 3.0. Griffith's argument has already stated that the local stress can exceed the remote tension by orders of magnitude at elliptical holes with a very large axial ratio.

Conversion of compression into tension

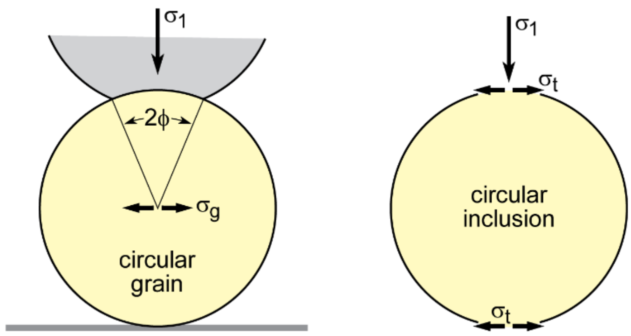
Experiments in compression have shown that flaws can induce local tensile stresses. For example, for a small angle 2ϕ of grain contact, stress σ_g at the grain center is tensile:

$$\sigma_g = -\sigma_{1r}\left(2\phi/\pi\right)$$

At the ends of the inclusion diameter parallel to the applied compression σ_{1r} the tangential stress is

$$\sigma_t = -\sigma_{1r}\left[(1-k)/(2k+1)\right]$$

In both cases, the change in sign shows that remote compression changes into a local tension. Since compressive stresses are large in the Earth's crust, and the tensile strength of rock is small, this conversion provides an attractive mechanism for joint initiation.



Conversion of compression into tensile stress on and in particles

Griffith's argument has also demonstrated that sliding of the walls of an elliptical crack inclined to the remote compression induces tensile stresses at its extremities (wing cracks at a larger scale).

Tension stresses due to pore pressure

Fluid pressures exceeding the least compressive remote stress produce tensile stresses greater than the rock tensile strength, in particular at the extremities of elliptical cavities and micro-cracks. Because the stress concentration increases with joint length (effect of aspect ratio), the tip propagation velocity increases as joints grow and can propagate as long as adequate fluid pressure is maintained (up to a limiting value).

Tip propagation

Fractures may lengthen in a direction parallel to their initial plane (**penetration** of the host material) or may deviate from the initial fracture orientation (**refraction** and **deflection**). Pre-existing discontinuities such as bedding surfaces may interrupt propagation. The former may be the site of

step-overs, where two parallel, not coplanar fractures were initially offset at the time of their formation.

Maximum depth of formation

Assuming that σ_1 is vertical near the surface of the Earth, it is the weight of the rocks lessened by the pore pressure P_f. The pore pressure is commonly expressed by the fluid pressure ratio $\lambda = P_f/\rho gz$. Hence σ_1 can be expressed as a function of the depth z and the rock density ρ as:

$$\sigma_1 = \rho gz(1-\lambda)$$

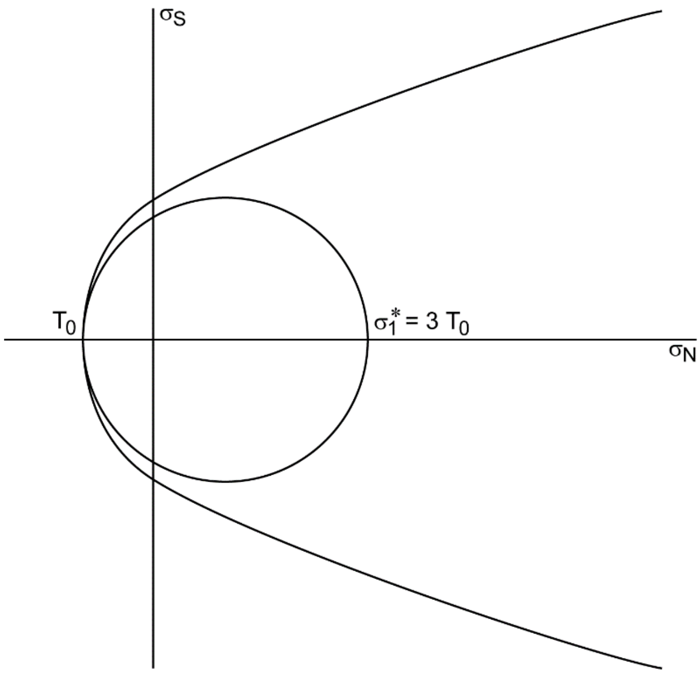
With this assumption, $\sigma_3 = T_0$ is horizontal and negative.
Considering the unique solution for a "pure" joint on a Mohr plot (i.e. the single point where the failure envelope cuts the σ_N axis, on its negative side) and that the differential stress is related to the tensile strength $(\sigma_1 - \sigma_3 > 3T_0)$, the maximum depth of formation of a tensile joint is:

$$z_{max} = \frac{3T_0}{\rho g(1-\lambda)}$$

Taking the standard value of $T_0 = 40$ MPa for rocks, the maximum depth of formation of joints is ca. 6 km. An exception occurs under high fluid pressure.

Exercise

Explain why one assumes that σ_1 is vertical near the surface of the Earth. Under such circumstances, σ_1 is the lithostatic pressure that may be reduced by pore pressure; justify this statement. Considering the unique solution for the nucleation of a tensile joint, and accepting that the tensile strength of a rock is 40 MPa, calculate the maximum depth at which jointing may occur.



Unloading joints

Rocks possess elastic properties closely related to those measured in the laboratory. In non-orogenic environments, uplift and exhumation give rise to changes in the horizontal and vertical stresses, which may exceed the tensile strength of rocks. In particular, hydrostatic confining pressure is released during decompression and the buried rocks tend to expand radially. However, the differences in **compressibility** between adjacent lithologies and mineral grains of different orientation or composition trigger local deviatoric stresses. These local, non-hydrostatic stress generated by decompression are such that the two principal stresses σ_2 and σ_3 may be tensile above a critical depth.

Sheet (exfoliation) joints

Erosion relieves vertical stress, which must approach one atmosphere, but lateral stress (at least the lithostatic pressure) is not reduced proportionally. Therefore, the state of stress becomes non-hydrostatic and the vertical stress becomes minimum principal stress σ_3 so that joints form approximately parallel to the earth's surface. Dilatant joints formed during erosion of homogeneous rocks such as granite are sub-parallel to the topography, and this orientation results in sets of flat-lying, curved and large joints referred to as **sheeting** or **sheet structure**. Spacing between sheet joints increases with depth, down to 50-100 m. Deeper sheet joints have a larger radius of curvature.
As soon as the stress is relieved in the vertical σ_3 direction, the original σ_2 becomes the greatest tensile stress. When the tensile strength is exceeded once more, a set of extension fractures perpendicular to the original set will form, generally somewhat less well developed than the first. The amount of expansion to be expected from the release of stored stress consequent on burial is indicated by the values of the **compressibility** of rocks, the ratio of volume change to pressure change.

Bedding-parallel and bedding-contained joints

Pressure changes of 200 MPa, corresponding with depth changes of about 6 km, lead to volume changes ranging from a few percents to a few tenths of a percent. If such volume changes are accomplished mainly by vertical extension, and if this extension takes place fast enough, horizontal dilatant joints may form. Parting of **bedding-parallel joints** is also related to unloading.
Decompression joints may also form vertically. They commonly abut against layer boundaries and dissect layered rocks in blocky elements. Such **bedding-contained joints** exhibit different spacing from bed to bed, which likely reflects differences in compressibility between different layers.
In homogeneous and isotropic rocks such as granite and sandstone, horizontal and vertical joints dissect rocks in near cubic elements. Weathering along these joints may lead to extreme rounding, which results in **boulders**.

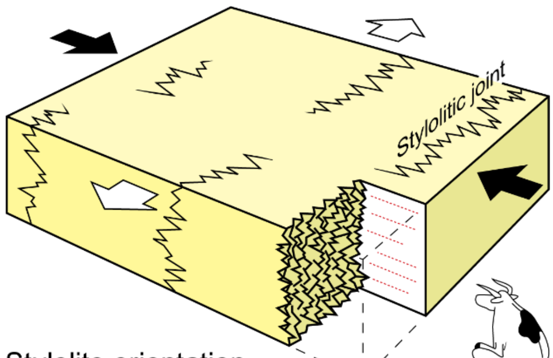
Joints due to non-decompressional volume changes.

On a regional scale, several rock types are juxtaposed in layers or other configurations. When these are cooled, local deviatoric stresses will be set up within them because of the differences in thermal contraction coefficients between different rock units and mineral grains in contact. Local non-hydrostatic stress generated by cooling is particularly important in joint formation in magmatic rocks. Regional deformation can impel material transfers that locally result in significant volume changes.

Stylolitic joints

Stylolitic joints have a characteristic saw-tooth profile and an interdigitating cone-like form in three dimensions. The interlocking ‘teeth’ are normal or oblique to the joint surface. Stylolitic joints are surfaces along which relatively soluble rock material has been removed by pressure-induced chemical dissolution to accommodate shortening. Shortening is parallel to the teeth direction. Stress concentration at the contact between grains triggers dissolution. Relatively insoluble residues (clay, iron oxides, etc) remained accumulated in the joint. This deformation mechanism is called **pressure solution**. Stylolites are particularly common in limestone.
Assuming that stylolitic joints initiate as flat planes and do not propagate out of plane, their amplitudes represent a minimum estimate of the amount of shortening (compaction) that has occurred. Assuming that insoluble material initially was evenly distributed in the rock and that there has been no contamination by circulating fluids, the thickness of insoluble residue along a stylolitic joint would be proportional to the amount of material dissolved and would, therefore, be proportional

to the shortening displacement across the stylolitic joint. Owing to this mode of displacements some authors call them **anticracks** or **mode 4 (closing mode) fractures**.

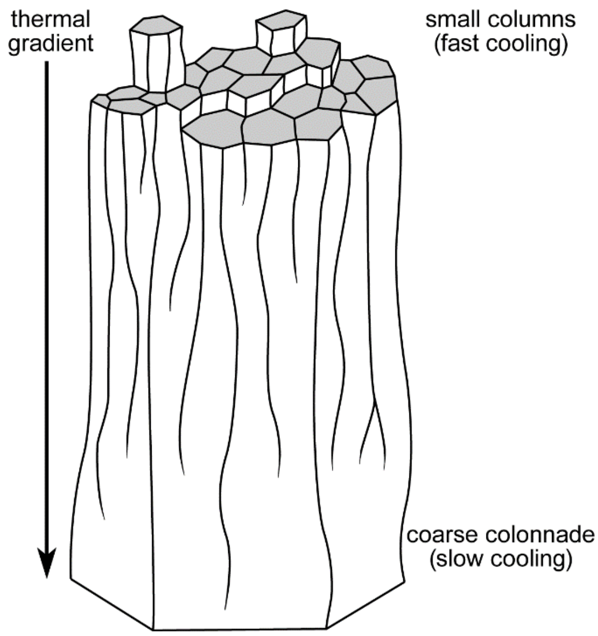


Stylolite orientation
with respect to compression

Columnar joints

Columnar joints are most prominent in basaltic sills and lava flows. They form a three-dimensional network of interconnected fractures that dissect the rock in long and spectacular polygonal (commonly five- or six-sided) columns. Minor, column-normal joint sets segment columns along their length and generally terminate at the column-bounding joints. Thermal contraction during cooling causes these column-bounding and column-normal joint sets to form and propagate perpendicular to isotherms.

No stress occurs if the temperature of a homogeneous, isotropic, and unconstrained body changes. Stresses arise if the body is prevented from expanding or contracting, or if there is an uneven temperature distribution, as it is the case from the cold top to the warm bottom of a lava flow. A joint begins to form when the local stresses are equal to or exceed the tensile strength of the rock (up to 485 MPa for basalt). Fracturing relieves thermal stresses along the joint sides, perpendicular to the fracture plane, but concentrates stresses at its tip. Column-bounding fractures propagate in the direction of the thermal gradient, following it as it moves through the cooling lava from the cool outside to the hotter lava. Thus, fractures grow by the successive addition of new segments to previous ones.



Sketched coarsening of a columnar joint pattern
adapted from Goehring & Morris 2005 *Europhys. Lett.* **69(5)** 739-745

Propagation occurs whenever the stress concentration at the fracture tip is greater than or equal to the tensile strength of the rock. Fracture growth ceases when insufficient thermal stress exists for propagation or when the strain rate is too low to overcome viscous relaxation of stress where the fracture tip is close to the cooling front, which is the transition between the cooled, brittle and the warm, viscous or visco-elastic lava.
Three joint patterns are common to many cooled lava flows in which isotherms (hence thermal stresses) are essentially horizontal. The lower part, called the **colonnade,** contains regular vertical columns. The central, approximately elliptical zone named **entablature**, often located slightly below the flow center, displays irregular twisted columns expressing rapid convective cooling. The upper zone may be regularly columnar **(upper colonnade)**, crudely columnar **(pseudo-columnar)**, or bulky. This distribution may reflect the fact that the conductive cooling rate controls joint spacing: fast cooling leads to narrow columns whose diameter increases inward from the margins of the flow.

Exercise

Two-dimensional thermal strain is linearly linked to the temperature change:

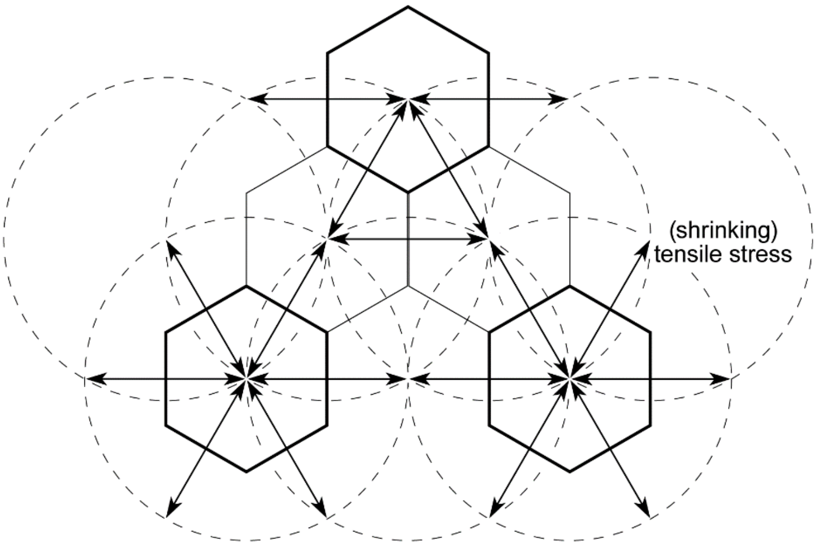
$$\varepsilon = \alpha \, \Delta T$$

where α *is the coefficient of thermal expansion in the considered reference frame. One assumes that strain is homogeneously distributed. Consider a basaltic lava flow erupted at 1200°C, which will cool down to 0°C (we are in Iceland). Openings along columnar joints will compensate thermal retraction of a 5000m long flow that cannot shrink because it adheres to its base. Calculate the linear strain with* $\alpha = 2.5\,10^{-6}\,°C^{-1}$. *Assuming the "width" of joint is* $5\,10^{-4}\,m$, *what would be the average size of the columns? Is it geologically relevant?*

Jointing also occurs in intrusive igneous rocks because they contract more than the cooler country rock. Like for lavas, thermal stresses arise because the thermal shrinkage of the intrusive rock is not free. Downward movement of the overlying country rock accommodates vertical contraction. If the boundary between the igneous rock and the country rocks remains coherent, then compression structures must develop in the country rock or extensional structures must develop in the cooling magma to accommodate its horizontal contraction.

Mudcracks

Desiccation-related **mud cracks** in sediments are polygonal patterns, ideally hexagonal, similar in many ways to columnar jointing of volcanic rocks. The incremental growth of desiccation joints is cinematically similar to cooling joints, but the nature of joint growth due to loss of water in layered rocks is not well known. The crack morphology is influenced by drying gradients with strong gradients having a strong orienting effect. The size of the polygons and the width and depth of the cracks depend on the thickness of the layer of wet mud.
Why hexagonal? This symmetry is a simple geometrical rule. If the rock is perfectly homogeneous and drying (or cooling) is also perfectly uniform, then the shrinkage centers are equally distributed. The distance between all centers is equal, which in planar view yields six circles tangent to each other and centered on one imbricate circle. Joints develop orthogonal to the tensile stresses, which are equal from center to center. Hexagons occur under these ideal conditions.



Development of a hexagonal joint pattern due to tensile pull towards equally distributed cooling or drying centers within a homogeneous material

Joints due to regional deformation.

Many joints, and particularly those that cut through different lithologies, are related directly to folds produced by regional deformation. The folds may be pronounced features or barely perceptible regional upwarps or downwarps. Indeed, a conceivable cause of regional jointing is the very gentle flexing of lithospheric plates to be expected when a plate changes latitude and thereby its radius of curvature. Joints geometrically related to folds may originate during the folding. In that case, they may reflect some elongation of the rocks, often parallel to fold hinge lines. If produced after folding, their orientation must be accounted for by the mechanical anisotropy of the folded rock.

Veins

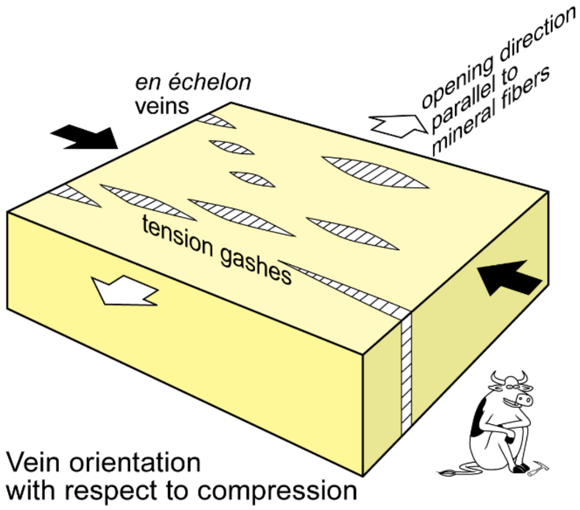
Veins are dilated fractures filled with oriented crystal fibers or non-oriented mineral deposits (typically quartz, calcite or carbonates). Such secondary crystallizations have been transported into and then deposited or precipitated along the fracture from solutions under favorable conditions of temperature and pressure. Veins are thus taken as evidence for the movement of fluids along fractures. They occur in rocks of all types and metamorphic grades with thickness from less than a millimeter to several meters. They have accommodated localized extension whose amount equals the space they occupy unless there was wall dissolution during early fluid circulation.
Non-fibrous, massive crystals generally grow in open cavities and the veins themselves, sometimes called **fault-cast veins,** may contain open gaps in which euhedral crystal terminations grow. Mineralized veins occasionally contain economically important concentrations of metals or other useful elements in **ore deposits**. Importantly also, they often contain datable material.
Veins are sealed fractures, hence evidence for reduced permeability. They also strengthen the rock and restore continuity across the fracture. If the vein is weaker than the host rock, repeated fracturing localizes in the vein or at the rock interface; conversely, renewed cracking tends to occur in the surrounding rock if the vein material is stronger than the rock. Each vein then reflects a single fracture event, which is known as **crack-jump** mechanism. Groups of veins constitute a **vein array**.

Tension veins

Description; definition

Tension gashes and veins are filled mode 1 fractures formed in response to combined tectonic and pore pressures. The latter have been high enough to enable a tensile effective minimum stress σ_3^{eff} normal to the plane of fracture. The cohesion of the vein-rock system determines where new fractures occur:

- Healed fractures restore the mechanical strength of systems in which cemented veins tend to remain sealed and new fractures occur in the host rock.
- Weak fractures localize subsequent rupture and grow at each cracking event.



Fiber minerals

Elongate, infilling minerals crystallize while the vein opens, with the long axis of the fibers tracking the incremental extension direction. Therefore, fiber minerals track the opening trajectory and document the formation of the tension fracture. The mineral nature of these fibers points to diagenetic or metamorphic conditions and fluid compositions.

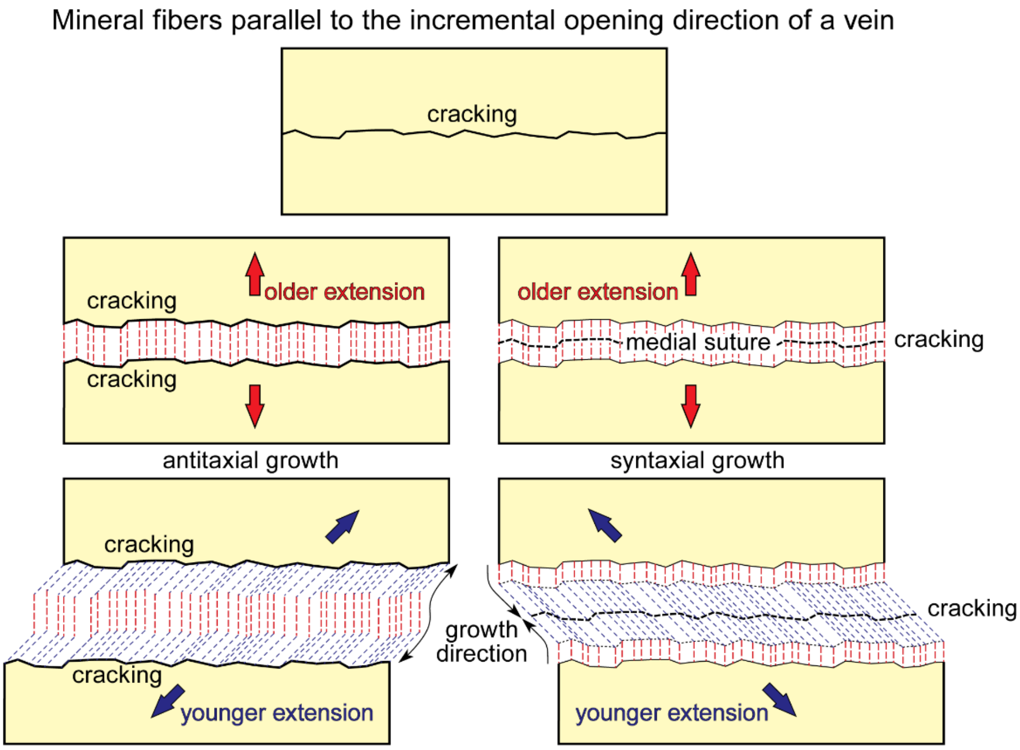
Three types of internal structure are used to relate the shape and orientation of veins with strain orientation:

- Undeformed fibers are perpendicular to the margin of purely extensional veins.
- Undeformed (straight) fibers in shear veins are oblique to the vein margin according to the shear component parallel to vein walls during vein opening.
- Fibrous crystals often grow at a high angle to the vein wall in a curved shape, without any bending of the lattice, indicating that the curved shape is a growth feature.

The originally curved fibers (they are not deformed) record a component of rotation during the opening direction of the fracture. They indicate that the opening direction changed during vein formation.

Three types of growth direction of vein crystals relative to wall rock are identified:

- **Syntaxial (inward) growth** adds material along the center of the vein. Fibers grow from the wall in optical continuity with mineral grains of the same composition in the host rock. Crystallization is interpreted as progressive growth from wall to center during vein opening, in the opening direction. Fibers extending from opposite walls meet at a **medial suture** (sometimes on one side) where there is both a structural and optical discontinuity. The medial plane is the fracture plane where fiber separation is continuously sealed by new material added to both sides of the medial suture during successive cracking. This indicates that the medial suture keeps having the weakest tensional strength of the vein-rock system.
- **Antitaxial (outward) growth** adds material along the vein wall. This occurs when the fiber mineral is absent or uncommon in the host rock. Single fibrous crystals, running from wall to wall, have grown outward from the median zone of the vein, which may contain wall rock inclusions. This implies that the two walls between the vein and the rock are two simultaneous fracture/growth planes where new material is continuously added. This feature suggests that adhesion along the rock-vein interface is weaker than the bulk tensional strength.



- **Ataxial growth** refers to veins due to repeated opening and sealing of fracture planes jumping positions within the growing vein. Fibers are stretched (microboudinaged) crystals. Such structures suggest that the vein material has the weakest tensional strength of the vein-rock system.

Crack-seal mechanism

Linear bands of regularly oriented and spaced solid and fluid inclusions parallel to the vein wall, across mineral fibers, suggest repeated microfracturing of the fibers (microboudinage) followed by deposition of optically continuous overgrowth that heals the fracture: the **crack-seal** mechanism. Growth preserves inclusions indicating repeated fracturing. Each microfracturing event records an incremental stage of vein opening.

Dikes

Veins filled by rock are classified as **dikes**.
Most commonly, dikes are magma-filled fractures.
There are also two types of sediment-filled fractures:
Intrusive clastic dikes contain fluidized or brecciated sediments that have been hydraulically injected into fractures of the overlying sediment. The injection is due to high pore pressure in the unconsolidated, source sediments. Clastic dikes commonly use joint patterns.
Neptunian dikes represent sediment filling, from above, of open fractures that may penetrate dip in the layers. Neptunian dikes also are commonly consistent with joint patterns.

Joints and veins in relations to other structures

Fracture sets often have consistent geometrical relationships to other structural orientations.

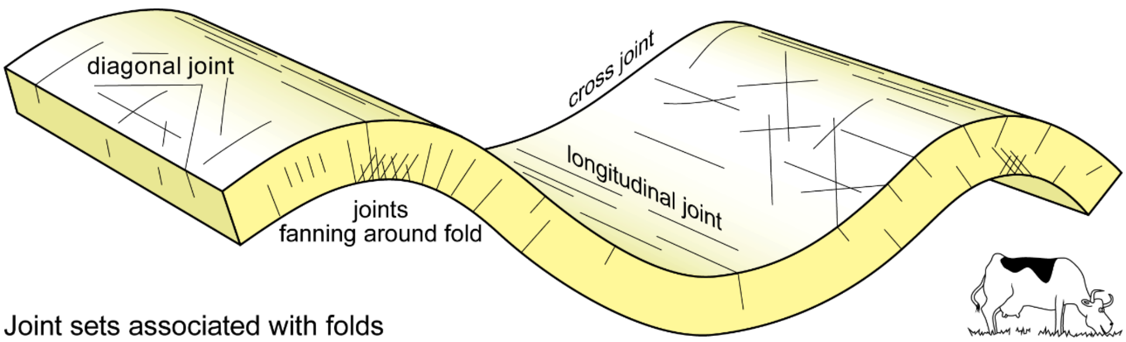
Tabular regions

In flat-lying sediments that have undergone little or no deformation, the most prominent joints usually are vertical and exhibit a marked consistency in their orientation patterns. The interpretation of these joint systems almost orthogonal to layers rarely leads to unequivocal conclusions regarding the stress or strain history of the area surveyed. However, such patterns are common in the forelands of mountain belts, suggesting that portions of the upper crust are subject to rectilinear stress fields.

Folded regions

Joints are often a part of the deformation in regions where rocks have been folded. Although joints normally are nearly perpendicular to bedding, they commonly form in a predictable pattern with respect to the hinge trends.

- **Longitudinal joints** are roughly parallel to fold axes and often fan around the fold.
- **Cross-joints** are approximately perpendicular to fold axes. They are common and indicate axis-parallel extension.
- **Diagonal joints** generally occur in paired, conjugate sets oblique to the fold axes, more or less symmetrically arranged about the longitudinal and cross-joints.
- **Strike joints** are parallel to the strike of fold axial planes, whereas **cross-strike joints** cut across the axial plane.



Joint sets associated with folds

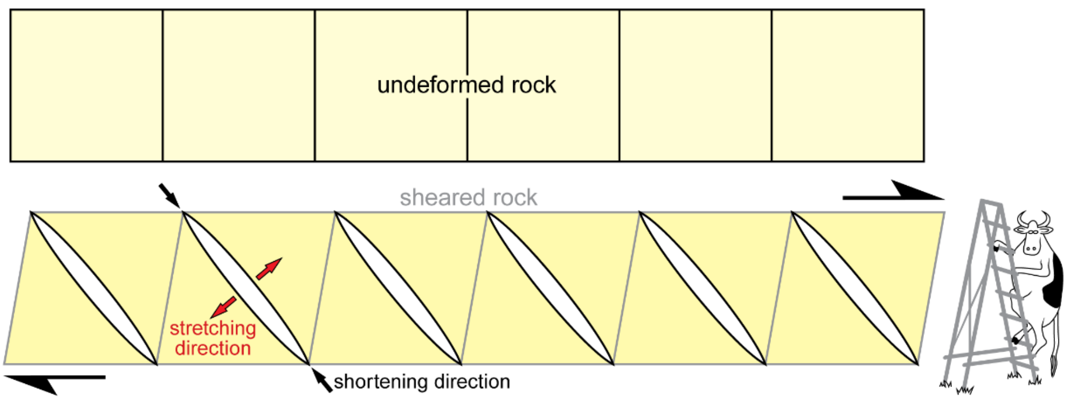
Gaussian Curvature or stress history analysis can predict the orientation and relative intensity of fracturing within folded structures. The Gaussian curvature is the product (K) of the principal curvatures (k_1 and k_2), which follow the trends of the principal strain axes (X and Z). Extensional fractures will lie parallel to one of the principal curvatures, dependent on the stress field that formed the fold. The intensity of fracturing is proportional to the degree of curvature (bending) of the strata.

Faulted regions

Joints associated with faults may predate the faults and therefore may be genetically unrelated to the faults apart from a geometrical control on the orientation of the fault planes. Joints adjacent to faults commonly occur with one set parallel to, and one set oblique to the fault planes, whatever the fault type. Joint density increases next to the fault.

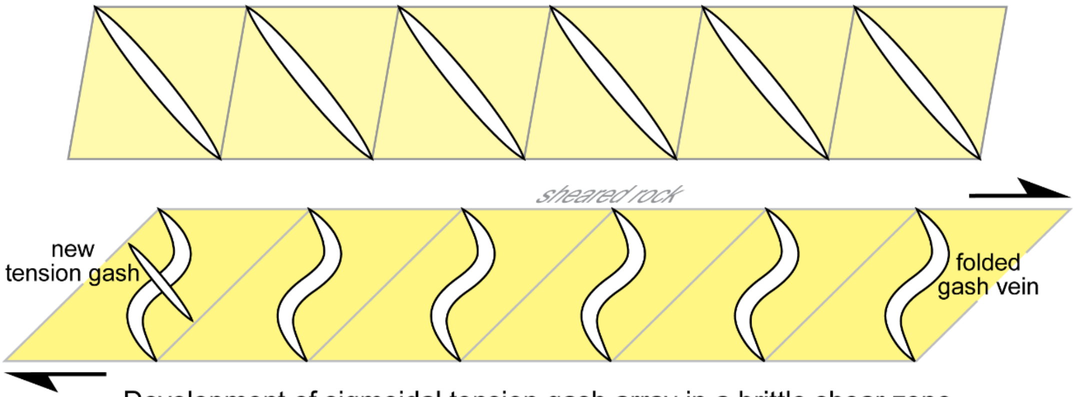
En échelon and sigmoidal veins

The tips of small, lens-shaped veins (**tension gashes**) propagate in a direction perpendicular to the incremental principal extension. *En échelon* veins are planar, regularly spaced and mutually parallel in an overlapping or staggered arrangement.



Development of *en échelon* tension gash array in a brittle shear zone
(instantaneous stretching and shortening directions do not rotate during shearing)

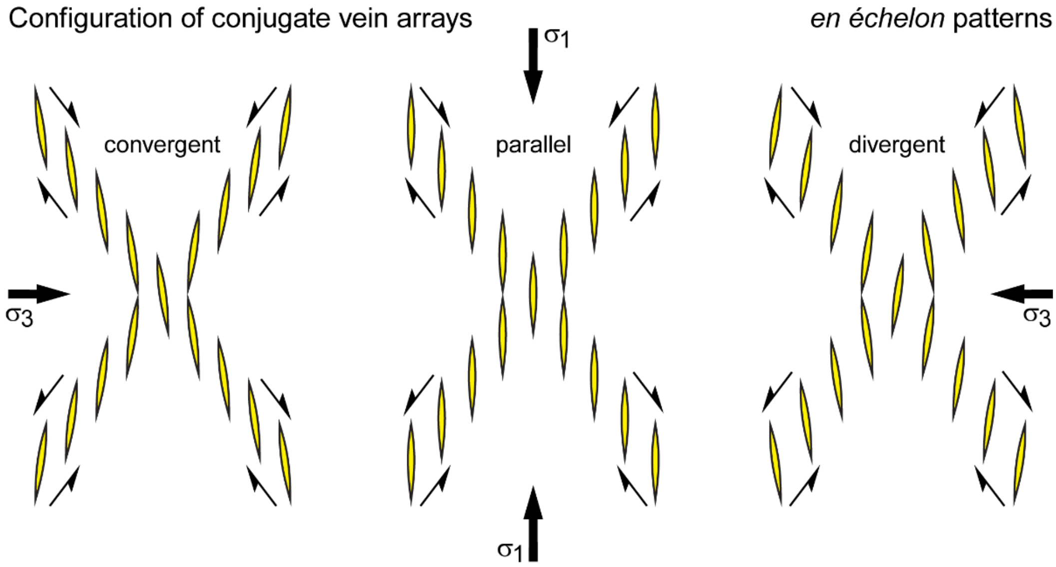
Each vein is relatively short but collectively they form a linear brittle shear zone delimited by two parallel, non-material enveloping surfaces. The strike of the individual veins is oblique to the linear zone as a whole, which is interpreted as a discrete, potential fault zone. *En échelon* veins are inclined against the sense of shear, in agreement with extension fractures being initiated normal to the incremental extension within the fault zone. Pressure solution seams suborthogonal to veins are common in the adjacent rock.Since tension gashes initiate as early increments of brittle strain, the alignment of *en échelon* veins is parallel to a **potential** shear fracture initially **segmented** in parallel and obliquely aligned dilatant joints or veins. Veins and associated fractures link and form a larger fault zone during further deformation.



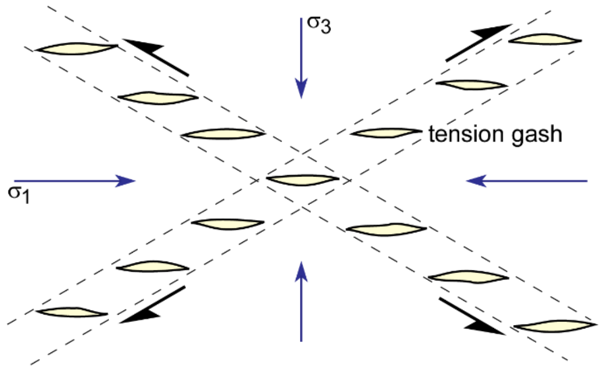
Development of sigmoidal tension gash array in a brittle shear zone
(instantaneous stretching and shortening directions do not rotate during shearing)

En échelon veins become **sigmoidal** (i.e. S- or Z-shaped) when the central part of the vein (along with the rock bridges in between) has rotated while the vein was lengthening during deformation. Any new increment of vein lengthening or opening of new veins tracks incremental principal strain directions (lengthening parallel to the far-field maximum compression, opening parallel to extension). The sense of rotation of the central part of the vein with respect to the tips indicates the sense of shear. Two geometrical types of conjugate arrays are identified:

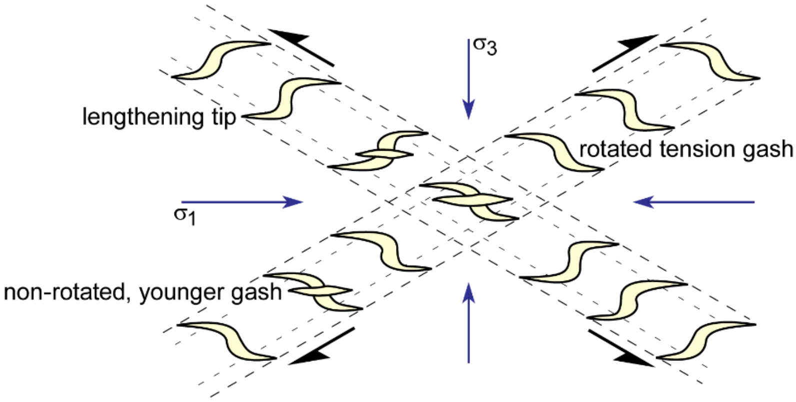
- Conjugate pairs of *en échelon* arrays where the undistorted portions of veins in both arrays are parallel; the length of the extension fractures bisects the acute angle between potential conjugate shear fractures.



- Conjugate pairs of *en échelon* arrays where the undistorted portions of veins in one array are not parallel to the undistorted portions of veins in the other array. In that case, the pair is divergent if the veins of both arrays diverge towards the intersection of the conjugate pair; conversely, the conjugate pair is convergent.



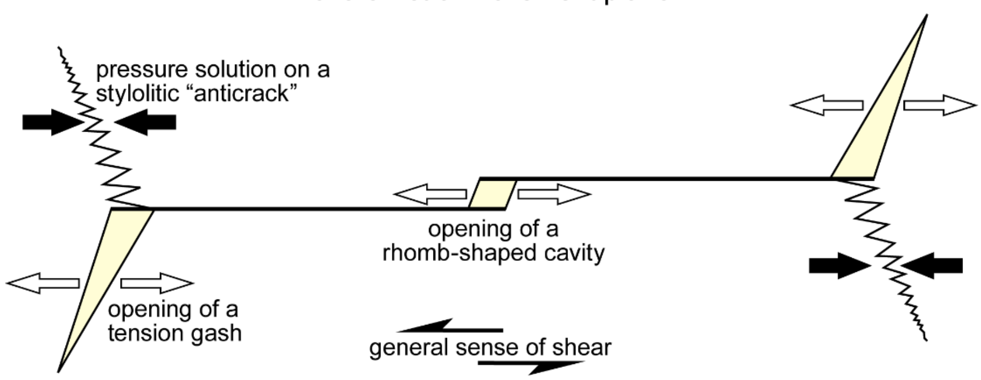
En échelon array of tension gashes along conjugate, brittle shear zones
and
sigmoidal gashes due to heterogeneous rotation of gashes in these shear zones



Tension gashes and "anticrack" stylolites

Vein boundaries move away from one another as the vein opens. Conversely, the surfaces of a stylolitic joint move towards each other as it forms, which originated the name of "anticrack". Paired **tension gashes** and stylolites may occur as pinnate fractures on opposite sides of small faults, in particular at the extremities of minor slip planes. The opening and closing directions help to define the slip direction.

Paired tension veins and stylolites at the tips
of a sinistral movement plane



Pinnate joints

The **pinnate** joints are arranged diagonally, often in an *en échelon* array in the immediate vicinity of the fault plane. They intersect the associated fault plane normal to the slip vector and subtend an acute angle with it that closes in the direction of relative movement of the blocks containing the joints. Depending on their orientation with respect to the relative movement of the fault block, they may be either shear or tension joints. Pinnate fractures can form both before and during slip on the associated fault. They have been produced in experiments on a wide variety of materials.

Intrusive bodies

Joint systems in igneous rocks can result from stresses arising during the cooling of the rock mass within or out of a regional stress field. Such joint systems may be quite different from joint systems in the surrounding rocks.

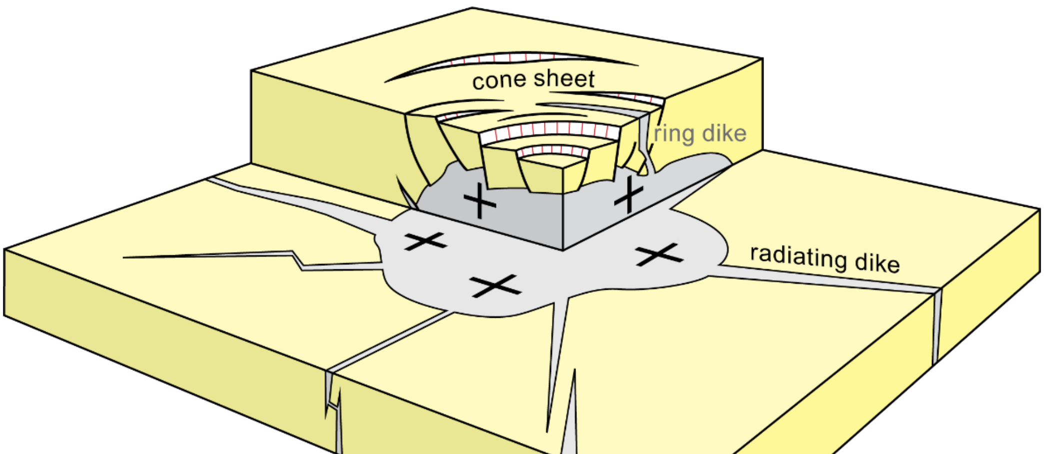
Primary fractures of plutonic rocks

Primary fracture systems (primary joint systems) of plutonic rocks are dilatant veins and dikes directly associated with the emplacement and with the fabric flow of plutonic bodies. They commonly are veins of igneous differentiates (aplite to pegmatite) and veins coated by hydrothermal and deuteric minerals. Four main systems have been identified and were awkwardly given the same names as joint systems in deformed sediments.

- Cross-joints are consistently normal to the flow line and plane; they are considered as standard extension fractures and many carry vein fillings (magmatic and mineral).
- Longitudinal joints (S-joints) are nearly always steeply dipping veins parallel to the flow lines and orthogonal to the flow planes.
- Diagonal joints (or marginal joints) form acute dihedral angles with the cross-joints. They are steep veins oblique to flow lines and orthogonal to oblique to flow planes.
- Flat-lying joints (or stretch joints) are parallel to the flow planes. They often are confined to the upper parts of intrusions.

Brittle joints parallel to primary joints are often symmetrically related to the contacts of the body, suggesting an origin of primary joints during emplacement and cooling.

Dikes habitually form sets and **swarms** associated with plutonic bodies. Injection along joints is their main mode of emplacement. Therefore, they serve to outline the failure pattern associated with intrusion. **Radiating** dikes are common around volcanic necks and shallow-depth (**hypabyssal**) intrusions. **Ring dikes** and **cone sheets** are concentric around an intrusive center.



Joints, dikes and faults
around a central intrusion with a caldera

Summary

Joints are planar and curved fractures along which no appreciable shear displacement has taken place. Surface marks show evidence for nucleation, very fast propagation and arrest of mode 1, opening fractures. Joint propagation paths twist and tilt as they grow. Joint sets form in sedimentary and crystalline rocks, early, intermediate and late in their histories. They form at shallow depth, under low confining pressure and low temperature. Deep veins may form under high fluid pressure. Joints and veins represent an elastic response of the rock to changing strain and stress conditions. For example, some joints form during erosion unloading, because of the greater ease with which decompressed rock expands normal to, rather than parallel to, the free surface. However, there is no general or single origin of joints.
At and near the Earth's surface, joints may be weathered to produce open channels to fluid circulation and cemented by secondary minerals.

Recommended literature

Beach A. - 1975. The geometry of en-échelon vein arrays. *Tectonophysics.* **28** **(4)**, 245-263, 10.1016/0040-1951(75)90040-2

Hancock P.L. - 1985. Brittle microtectonics: principles and practice. *Journal of Structural Geology.* **7** **(3-4)**, 437-457, 10.1016/0191-8141(85)90048-3

Hodgson R.A. - 1961. Classification of structures on joint surfaces. *American Journal of Science.* **259** **(7)**, 493-502, 10.2475/ajs.259.7.493

Pollard D.D. & Fletcher R.C. - 2005. *Fundamentals of structural geology*. Cambridge University Press, Cambridge. 500 p.

Ramsay J.G. & Huber M.I. - 1987. *The techniques of modern structural geology - Volume2 : Folds and fractures*. Academic Press, London. 700 p.

Twiss R.J. & Moores E.M. - 1992. *Structural geology*. W.H. Freeman & Company, New York. 532 p.