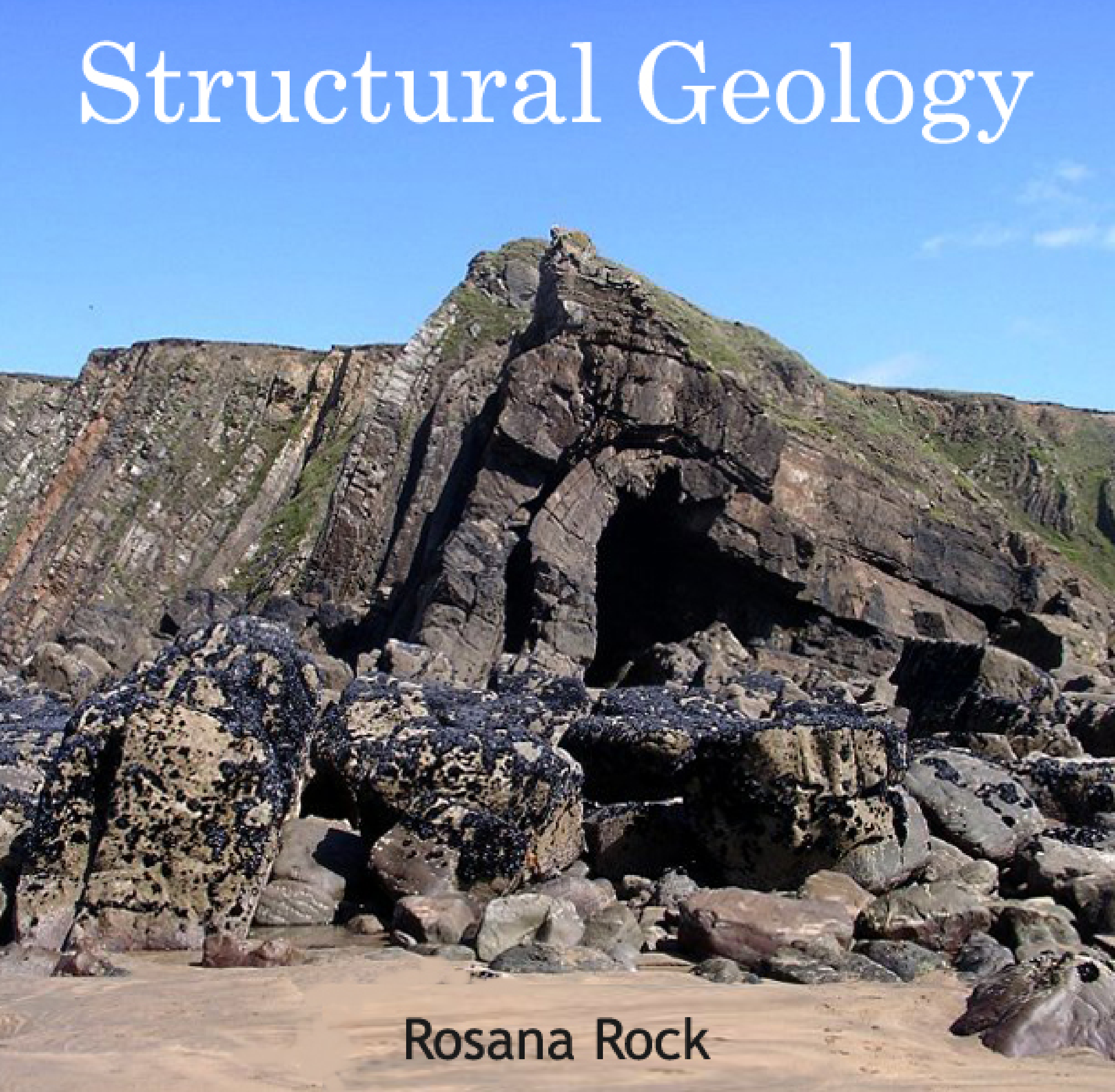
Structural Geology



Rosana Rock

First Edition, 2012

ISBN 978-81-323-1072-3

Published by:
College Publishing House
4735/22 Prakashdeep Bldg,
Ansari Road, Darya Ganj,
Delhi - 110002
Email: info@wtbooks.com

Table of Contents

Chapter- 1

Introduction to Structural Geology

Structural geology is the study of the three-dimensional distribution of rock units with respect to their deformational histories. The primary goal of structural geology is to use measurements of present-day rock geometries to uncover information about the history of deformation (strain) in the rocks, and ultimately, to understand the stress field that resulted in the observed strain and geometries. This understanding of the dynamics of the stress field can be linked to important events in the regional geologic past; a common goal is to understand the structural evolution of a particular area with respect to regionally widespread patterns of rock deformation (e.g., mountain building, rifting) due to plate tectonics.

Use and importance

The study of geologic structures has been of prime importance in economic geology, both petroleum geology and mining geology. Folded and faulted rock strata commonly form traps for the accumulation and concentration of fluids such as petroleum and natural gas. Faulted and structurally complex areas are notable as permeable zones for hydrothermal fluids and the resulting concentration areas for base and precious metal ore deposits. Veins of minerals containing various metals commonly occupy faults and fractures in structurally complex areas. These structurally fractured and faulted zones often occur in association with intrusive igneous rocks. They often also occur around geologic reef complexes and collapse features such as ancient sinkholes. Deposits of gold, silver, copper, lead, zinc, and other metals, are commonly located in structurally complex areas.

Structural geology is a critical part of engineering geology, which is concerned with the physical and mechanical properties of natural rocks. Structural fabrics and defects such as faults, folds, foliations and joints are internal weaknesses of rocks which may affect the stability of human engineered structures such as dams, road cuts, open pit mines and underground mines or road tunnels.

Geotechnical risk, including earthquake risk can only be investigated by inspecting a combination of structural geology and geomorphology. In addition areas of karst landscapes which are underlain by underground caverns and potential sinkholes or collapse features are of importance for these scientists. In addition, areas of steep slopes are potential collapse or landslide hazards.

Environmental geologists and hydrogeologists or hydrologists need to understand structural geology because structures are sites of groundwater flow and penetration, which may affect, for instance, seepage of toxic substances from waste dumps, or seepage of salty water into aquifers.

Plate tectonics is a theory developed during the 1960s which describes the movement of continents by way of the separation and collision of crustal plates. It is in a sense structural geology on a planet scale, and is used throughout structural geology as a framework to analyze and understand global, regional, and local scale features.

Methods

Structural geologists use a variety of methods to (first) measure rock geometries, (second) reconstruct their deformational histories, and (third) calculate the stress field that resulted in that deformation.

Geometries

Primary data sets for structural geology are collected in the field. Structural geologists measure a variety of planar features (bedding planes, foliation planes, fold axial planes, fault planes, and joints), and linear features (stretching lineations, in which minerals are ductily extended; fold axes; and intersection lineations, the trace of a planar feature on another planar surface).

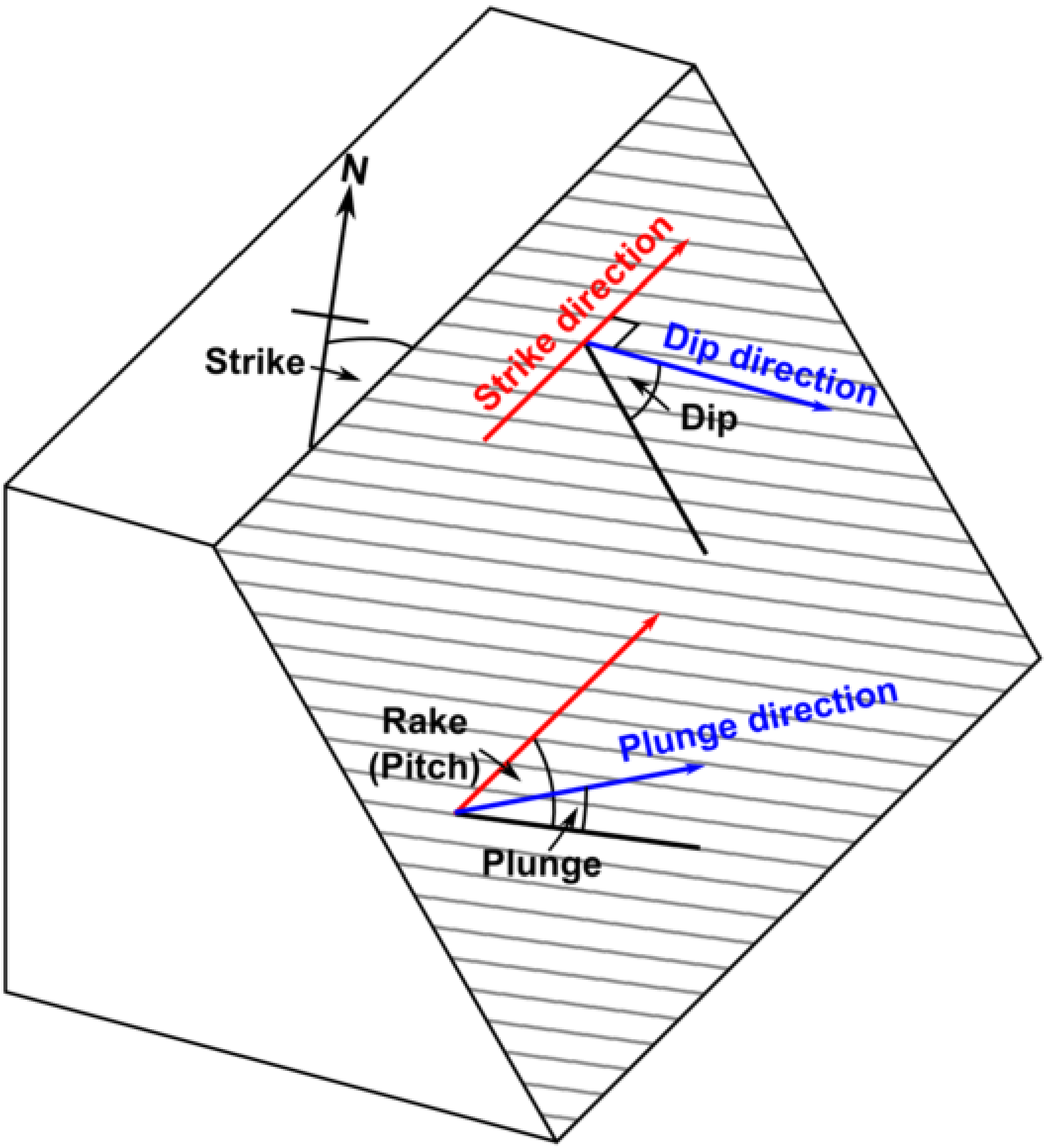


Illustration of measurement conventions for planar and linear structures

Measurement conventions

The inclination of a planar structure in geology is measured by *strike and dip*. The strike is the line of intersection between the planar feature and a horizontal plane, taken according to the right hand convention, and the dip is the magnitude of the inclination, below horizontal, at right angles to strike. For example; striking 25 degrees East of North, dipping 45 degrees Southeast, recorded as N25E,45SE.

Alternatively, dip and dip direction may be used as this is absolute. Dip direction is measured in 360 degrees, generally clockwise from North. For example, a dip of 45 degrees towards 115 degrees azimuth, recorded as 45/115. Note that this is the same as above.

The term *hade* is occasionally used and is the deviation of a plane from vertical i.e. (90°-dip).

Fold axis plunge is measured in dip and dip direction (strictly, plunge and azimuth of plunge). The orientation of a fold axial plane is measured in strike and dip or dip and dip direction.

Lineations are measured in terms of dip and dip direction, if possible. Often lineations occur expressed on a planar surface and can be difficult to measure directly. In this case, the lineation may be measured from the horizontal as a *rake* or *pitch* upon the surface.

Rake is measured by placing a protractor flat on the planar surface, with the flat edge horizontal and measuring the angle of the lineation clockwise from horizontal. The orientation of the lineation can then be calculated from the rake and strike-dip information of the plane it was measured from, using a stereographic projection.

If a fault has lineations formed by movement on the plane, eg; slickensides, this is recorded as a lineation, with a rake, and annotated as to the indication of throw on the fault.

Generally it is easier to record strike and dip information of planar structures in dip/dip direction format as this will match all the other structural information you may be recording about folds, lineations, etc., although there is an advantage to using different formats that discriminate between planar and linear data.

Plane, fabric, fold and deformation conventions

The convention for analysing structural geology is to identify the **planar structures**, often called *planar fabrics* because this implies a textural formation, the **linear structures** and, from analysis of these, unravel **deformations**.

Planar structures are named according to their order of formation, with original sedimentary layering the lowest at S0. Often it is impossible to identify S0 in highly deformed rocks, so numbering may be started at an arbitrary number or given a letter (S_A, for instance). In cases where there is a bedding-plane foliation caused by burial metamorphism or diagenesis this may be enumerated as S0a.

If there are folds, these are numbered as F_1, F_2, etc. Generally the axial plane foliation or cleavage of a fold is created during folding, and the number convention should match. For example, an F_2 fold should have an S_2 axial foliation.

Deformations are numbered according to their order of formation with the letter D denoting a deformation event. For example D_1, D_2, D_3. Folds and foliations, because they are formed by deformation events, should correlate with these events. For example an F_2 fold, with an S_2 axial plane foliation would be the result of a D_2 deformation.

Metamorphic events may span multiple deformations. Sometimes it is useful to identify them similarly to the structural features for which they are responsible, eg; M_2. This may be possible by observing porphyroblast formation in cleavages of known deformation age, by identifying metamorphic mineral assemblages created by different events, or via geochronology.

Intersection lineations in rocks, as they are the product of the intersection of two planar structures, are named according to the two planar structures from which they are formed. For instance, the intersection lineation of a S_1 cleavage and bedding is the $L_{1\text{-}0}$ intersection lineation (also known as the cleavage-bedding lineation).

Stretching lineations may be difficult to quantify, especially in highly stretched ductile rocks where minimal foliation information is preserved. Where possible, when correlated with deformations (as few are formed in folds, and many are not strictly associated with planar foliations), they may be identified similar to planar surfaces and folds, eg; L_1, L_2. For convenience some geologists prefer to annotate them with a subscript S, for example L_{s1} to differentiate them from intersection lineations, though this is generally redundant.

Kinematics

Geologists use their measurements of rock geometries to understand histories of strain in the rocks. Strain can take the form of brittle faulting and ductile folding and shearing. Brittle deformation takes place in the shallow crust, and ductile deformation takes place in the deeper crust, where temperatures and pressures are higher.

Stress Fields

By understanding the constitutive relationships between stress and strain in rocks, geologists can translate the observed patterns of rock deformation into a stress field during the geologic past. The following list of features are typically used to determine stress fields from deformational structures.

- In perfectly brittle rocks, faulting occurs at 30° to the greatest compressional stress. (Byerlee's Law)
- The greatest compressive stress is normal to fold axial planes.

Chapter- 2

Rock Microstructure

Rock microstructure includes the texture of a rock and the small scale rock structures. The words "texture" and "microstructure" are interchangeable, with the latter preferred in modern geological literature. However, texture is still acceptable because it is a useful means of identifying the origin of rocks, how they formed, and their appearance.

Textures are *penetrative fabrics* of rocks; they occur throughout the entirety of the rock mass on a microscopic, hand specimen and often on an outcrop scale. This is similar in many ways to foliations, except a texture does not necessarily carry structural information in terms of deformation events and orientation information. Structures occur on hand-specimen scale and above.

Microstructure analysis describes the textural features of the rock, and can provide information on the conditions of formation, petrogenesis, and subsequent deformation, folding or alteration events.

Sedimentary microstructures

Description of sedimentary rock microstructure aims to provide information on the conditions of deposition of the sediment, the paleo-environment, and the provenance of the sedimentary material.

Methods involve description of clast size, sorting, composition, rounding or angularity, sphericity and description of the matrix. Sedimentary microstructures, specifically, may include microscopic analogs of larger sedimentary structural features such as cross-bedding, syn-sedimentary faults, sediment slumping, cross-stratification, etc.

Maturity

The maturity of a sediment is related not only to the sorting (mean grain size and deviations), but also to the fragment sphericity, rounding and composition. Quartz-only sands are more mature than arkose or greywacke.

Fragment shape

Fragment shape gives information on the length of sediment transport. The more rounded the clasts, the more water-worn they are. Particle shape includes form and rounding. Form indicates whether a grain is more equant (round, spherical) or platy (flat, disc-like, oblate); as well as sphericity.

Roundness

Roundness refers to the degree of sharpness of the corners and edges of a grain. The surface texture of grains may be polished, frosted, or marked by small pits and scratches. This information can usually be seen best under a binocular microscope, not in a thin section.

Composition

Composition of the clasts can give clues as to the derivation of a rock's sediments. For instance, volcanic fragments, fragments of cherts, well-rounded sands all imply different sources.

Matrix and cement

The matrix of a sedimentary rock and the mineral cement (if any) holding it together are all diagnostic.

Diagenetic features

Usually diagenesis results in a weak bedding-plane foliation. Other effects can include flattening of grains, pressure dissolution and sub-grain deformation. Mineralogical changes may include zeolite or other authigenic minerals forming in low-grade metamorphic conditions.

Sorting

Sorting is used to describe the uniformity of grain sizes within a sedimentary rock. Understanding sorting is critical to making inferences on the degree of maturity and length of transport of a sediment. Sorting can be expressed mathematically by the standard deviation of the grain-size frequency curve of a sediment sample, expressed as values of φ (phi). Values range from $<0.35\varphi$ (very well sorted) to $>4.00\varphi$ (extremely poorly sorted).

Metamorphic microstructure

The study of metamorphic rock microstructures aims to determine the timing, sequence and conditions of deformations, mineral growth and overprinting of subsequent deformation events.

Metamorphic microstructures include textures formed by the development of foliation and overprinting of foliations causing crenulations. The relationship of porphyroblasts to the foliations and to other porphyroblasts can provide information on the order of formation of metamorphic assemblages or facies of minerals.

Shear textures are particularly suited to analysis by microstructural investigations, especially in mylonites and other highly disturbed and deformed rocks.

Foliations and crenulations

On the thin section and hand specimen scale a metamorphic rock may manifest a linear penetrative fabric called a foliation or a cleavage. Several foliations may be present in a rock, giving rise to a crenulation.

Identifying a foliation and its orientation is the first step in analysis of foliated metamorphic rocks. Gaining information on when the foliation formed is essential to reconstructing a P-T-t (pressure, temperature, time) path for a rock, as the relationship of a foliation to porphyroblasts is diagnostic of when the foliation formed, and the P-T conditions which existed at that time.

Ductile shear microstructures

Very distinctive textures form as a consequence of ductile shear. The microstructures of ductile shear zones are S-planes, C-planes and C' planes. S-planes or *schistosity* planes are parallel with the shear direction and are generally defined by micas or platy minerals. Define the flattened long-axis of the strain ellipse. C-planes or *cissalement* planes form oblique to the shear plane. The angle between the C and S planes is always acute, and defines the shear sense. Generally, the lower the C-S angle the greater the strain. The C' planes are rarely observed except in ultradeformed mylonites, and form nearly perpendicular to the S-plane.

Other microstructures which can give sense of shear include

- sigmoidal veins
- mica fish
- rotated porphyroblasts

Igneous microstructure

Analysis of igneous rock microstructure may complement descriptions on the hand specimen and outcrop scale. This is especially vital for describing phenocrysts and fragmental textures of tuffs, as often relationships between magma and phenocryst morphology are critical for analysing cooling, fractional crystallization and emplacement.

Analysis of intrusive rock microstructures can provide information on source and genesis, including contamination of igneous rocks by wall rocks and identifying crystals which may have been accumulated or dropped out of the melt. This is especially critical for komatiite lavas and ultramafic intrusive rocks.

General principles of igneous microstructure

Igneous microstructure is a combination of cooling rate, nucleation rate, eruption (if a lava), magma composition and its relationships to what minerals will nucleate, as well as physical effects of wall rocks, contamination and especially vapor.

Grain texture

According to the texture of the grains, igneous rocks may be classified as

- pegmatitic: very large crystals
- phaneritic: rocks contain minerals with crystals visible to the unaided eye, commonly intrusive
- aphanitic: rapid cooling, crystal nucleation and growth is stunted, forming a uniform, fine grained rock
- porphyritic: containing phenocrysts in a fine groundmass
- vesicular: contains voids caused by trapped gas while cooling
- vitreous: glassy or hyaline without crystals
- pyroclastic: rock contain fragments of crystals, phenocrysts and rock fragments
- equigranular: rock crystals are all the same size

Crystal shapes

Crystal shape is also an important factor in the texture of an igneous rock. Crystals may be euhedral, subeuhedral or anhedral:

- *Euhedral* or *automorphic*, if the crystallographic shape is preserved.
- *Subeuhedral* or *Subhedral*, if only part is preserved.
- *Anhedral* or xenomorphic, if the crystals present no recognizable crystallographic forms.

Rocks composed entirely of euhedral crystals are termed *panidiomorphic*, and rocks composed entirely of subhedral crystals are termed *subidiomorphic*.

Porphyritic structure

Porphyritic structure is caused by the nucleation of crystal sites and the growth of crystals in a liquid magma. Often a magma can only grow one mineral at a time especially if it is cooling slowly. This is why most igneous rocks have only one type of phenocryst mineral. Rhythmic cumulate layers in ultramafic intrusions are a result of uninterrupted slow cooling.

When a rock cools too quickly the liquid freezes into a solid glass, or crystalline groundmass. Often vapor loss from a magma chamber will cause a porphyritic texture.

Embayments or 'corroded' margins to phenocrysts infer that they were being resorbed by the magma and may imply addition of fresh, hotter magma. Ostwald ripening is also used to explain some porphyritic igneous textures, especially orthoclase megacrystic granites.

Phenocryst shape: implications

A crystal growing in a magma adopts a habit which best reflects its environment and cooling rate. The usual phenocryst habit is the ones commonly observed. This may imply a 'normal' cooling rate.

Abnormal cooling rates occur in supercooled magmas, particularly komatiite lavas. Here, low nucleation rates due to superfluidity prevent nucleation until the liquid is well below the mineral growth curve. Growth then occurs at extreme rates, favoring slender, long crystals. Additionally, at crystal vertices and terminations, spikes and skeletal shapes may form because nucleation favors crystal edges. Spinifex or dendritic texture is an example of this result. Hence, the shape of phenocrysts can provide valuable information on cooling rate and initial magma temperature.

Spherulites

Spherulitic texture is the result of cooling and nucleation of material in a magma which has achieved supersaturation in the crystal component. Thus it is often a subsolidus process in supercooler felsic rocks. Often, two minerals will grow together in the spherulite. Axiolitic texture results from spherulitic growth along fractures in volcanic glass, often from invasion of water.

Graphic and other intergrowth textures

Intergrowths of two or more minerals can form in a variety of ways, and interpretations of the intergrowths can be critical in understanding both magmatic and cooling histories of igneous rocks. A few of the many important textures are presented here as examples.

Graphic, micrographic texture, and granophyric textures are examples of intergrowths formed during magmatic crystallization. They are angular intergrowths of quartz and alkali feldspar. When well-developed, the intergrowths may resemble ancient cuneiform writing, hence the name. These intergrowths are typical of pegmatite and granophyre, and they have been interpreted as documenting simultaneous crystallization of the intergrown minerals in the presence of a silicate melt together with a water-rich phase.

Intergrowths that form by exsolution are aids in interpreting cooling histories of rocks. Perthite is an intergrowth of K-feldspar with albite feldspar, formed by exsolution from an alkali feldspar of intermediate composition: the coarseness of perthitic intergrowths is related to cooling rate. Perthite is typical of many granites. Myrmekite is a microscopic, vermicular (worm-like) intergrowth of quartz and sodium-rich plagioclase common in granite; myrmekite may form as alkali feldspar breaks down by exsolution and silicon is transported by fluids in cooling rocks.

Iron-titanium oxides are extremely important, as they carry the predominant magnetic signatures of many rocks, and so they have played a major role in our understanding of plate tectonics. These oxides commonly have complex textures related both to exsolution and oxidation. For instance, ulvospinel in igneous rocks such as basalt and gabbro commonly oxidizes during subsolidus cooling to produce regular intergrowths of magnetite and ilmenite. The process can determine what magnetic record is inherited by the rock.

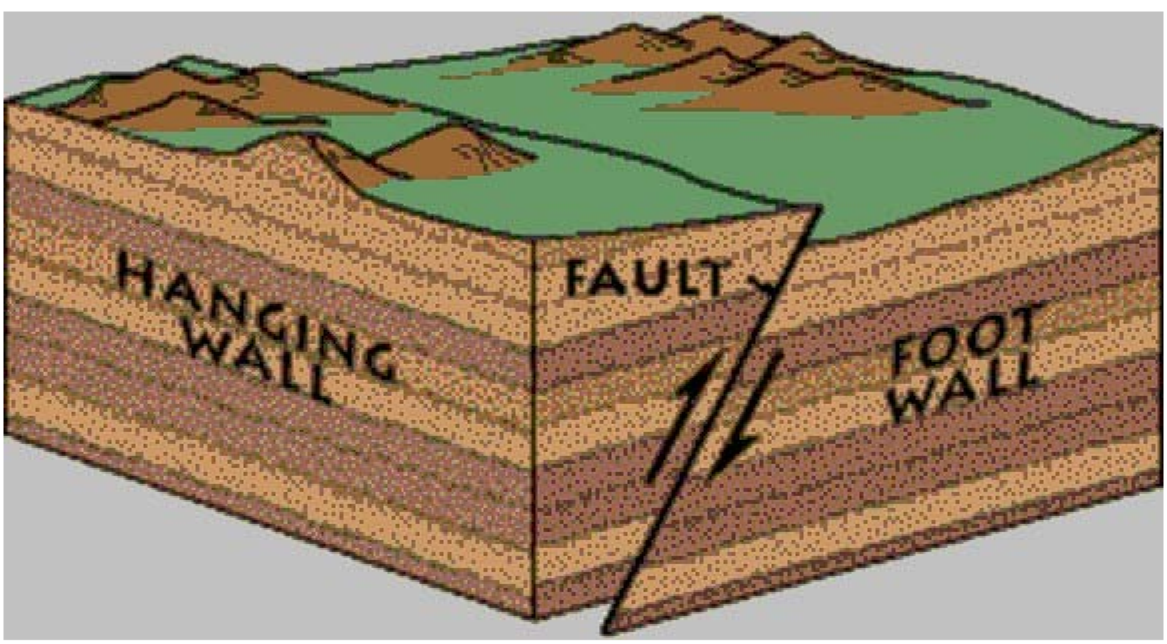
Chapter- 3

Fault and Fold

Fault

Minor normal faults in Triassic to Lower Jurassic Blomidon Formation sediments near Parrsboro, Nova Scotia

Ramon Fault on the southern side of Makhtesh Ramon, Israel



Hanging wall and footwall

In geology, a **fault** is a planar fracture or discontinuity in a volume of rock, across which there has been significant displacement. Large faults within the Earth's crust result from the action of tectonic forces. Energy release associated with rapid movement on active faults is the cause of most earthquakes.

A **fault line** is the surface trace of a fault, the line of intersection between the fault plane and the Earth's surface.

Since faults do not usually consist of a single, clean fracture, geologists use the term ***fault zone*** when referring to the zone of complex deformation associated with the fault plane.

The two sides of a non-vertical fault are known as the *hanging wall* and *footwall*. By definition, the hanging wall occurs above the fault and the footwall occurs below the fault. This terminology comes from mining: when working a tabular ore body, the miner stood with the footwall under his feet and with the hanging wall hanging above him.

Mechanics

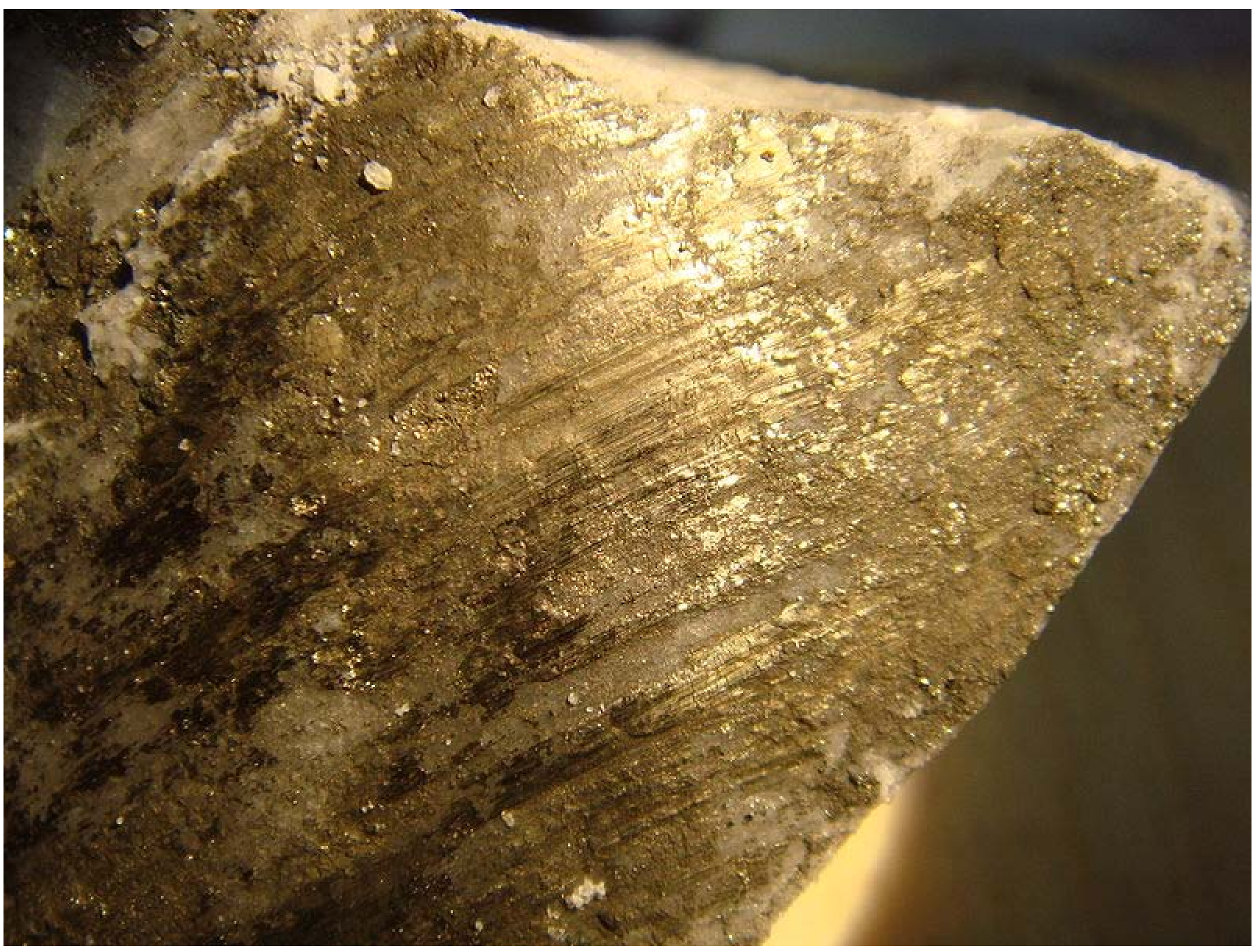
The Junction fault, dividing the Allegheny Plateau and the true Appalachian Mountains in Pennsylvania, United States

The relative motion of rocks on either side of the fault surface controls the origin and behavior of faults, in both an individual small fault and within the greater fault zones which define the tectonic plates.

Because of friction and the rigidity of the rock, the rocks cannot simply glide or flow past each other. Rather, stress builds up in rocks and when it reaches a level that exceeds the strain threshold, the accumulated potential energy is released as strain, which is focused into a plane along which relative motion is accommodated—the fault.

Strain is both accumulative and instantaneous depending on the rheology of the rock; the ductile lower crust and mantle accumulates deformation gradually via shearing, whereas the brittle upper crust reacts by fracture - instantaneous stress release - to cause motion along the fault. A fault in ductile rocks can also release instantaneously when the strain rate is too great. The energy released by instantaneous strain release causes earthquakes, a common phenomenon along transform boundaries.

Microfracturing and AMR theory



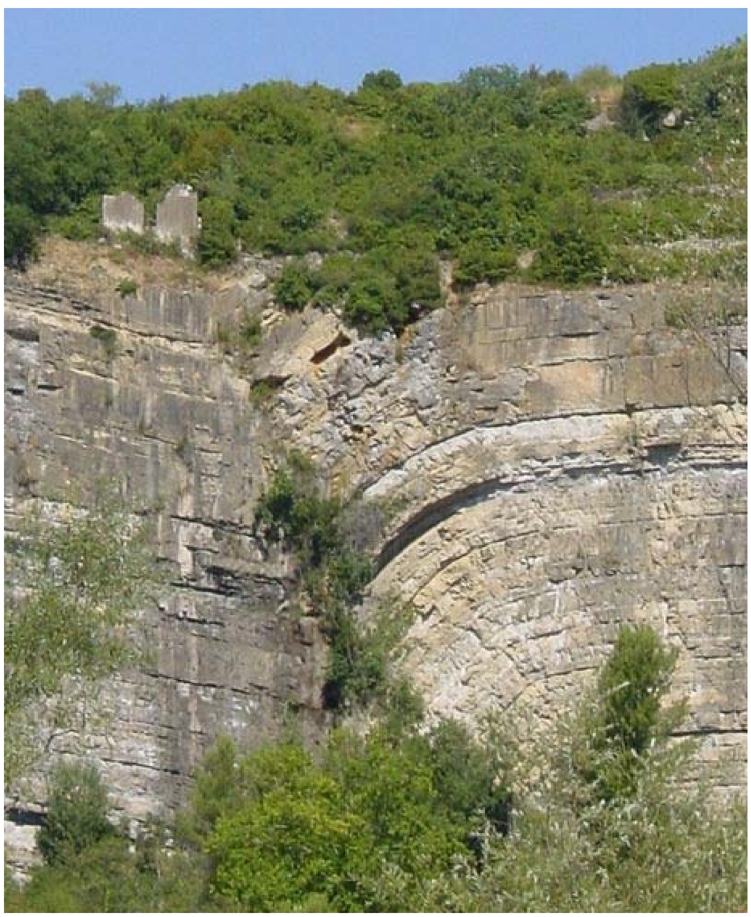
Dextral slickenside of pyrite on a possible microfault

Microfracturing, or microseismicity, is sometimes thought of as a symptom caused by rocks under strain, where small-scale failures, perhaps on areas the size of a dinner plate or a small area, release stress under high strain conditions. Only when sufficient microfractures link up into a large slip surface can a large seismic event or earthquake occur.

According to this theory, after a large earthquake, the majority of the stress is released and the frequency of microfracturing is exponentially lower. A related theory, accelerating moment release (**AMR**), hypothesizes that the seismicity rate accelerates in a well-behaved way prior to large earthquakes, and that it may provide a promising tool for earthquake prediction on the scale of days to years.

AMR is being increasingly used to predict rock failures within mines, and applications are being attempted for the portions of faults within brittle rheological conditions. Researchers observe similar behaviour in the tremors preceding volcanic eruptions.

Slip, heave, throw



A fault in the Grands Causses as seen from Bédarieux, France. The left side moves down while the right side moves up. The warping of the rock layers on the right is likely due to drag folding.

Slip is defined as the relative movement of geological features present on either side of a fault plane, and is a displacement vector. A fault's *sense of slip* is defined as the relative motion of the rock on each side of the fault with respect to the other side. In measuring the horizontal or vertical separation, the *throw* of the fault is the vertical component of the dip separation and the *heave* of the fault is the horizontal component, as in "throw up and heave out".

The vector of slip can be qualitatively assessed by studying the fault bend folding, i.e., the drag folding of strata on either side of the fault; the direction and magnitude of heave and throw can be measured only by finding common intersection points on either side of the fault. In practice, it is usually only possible to find the slip direction of faults, and an approximation of the heave and throw vector.

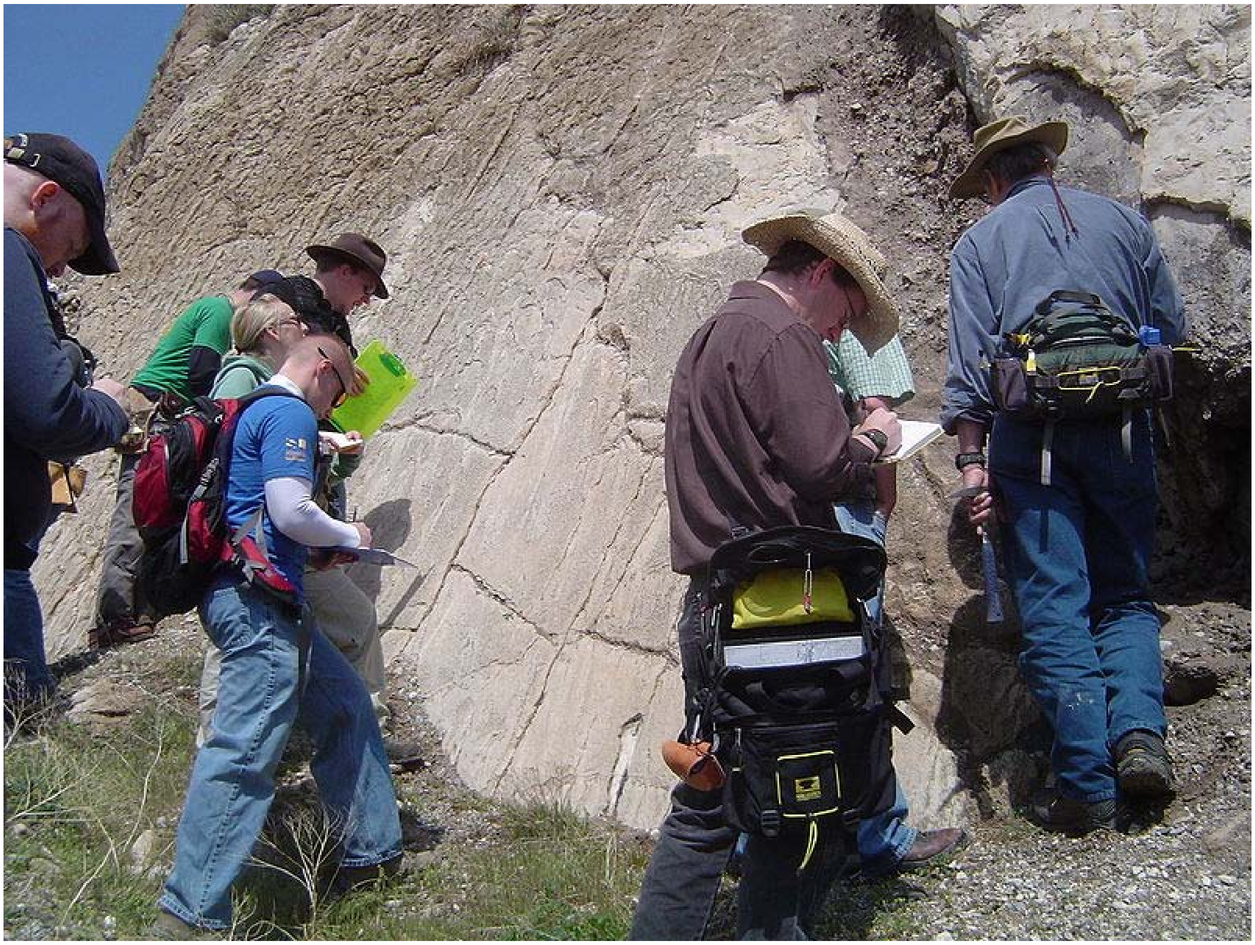
Fault types

Geologists can categorize faults into three groups based on the sense of slip:

1. a fault where the relative movement (or slip) on the fault plane is approximately vertical is known as a dip-slip fault
2. where the slip is approximately horizontal, the fault is known as a transcurrent or strike-slip fault
3. an oblique-slip fault has non-zero components of both strike and dip slip.

For all naming distinctions, it is the orientation of the **net** dip and sense of slip of the fault which must be considered, not the present-day orientation, which may have been altered by local or regional folding or tilting.

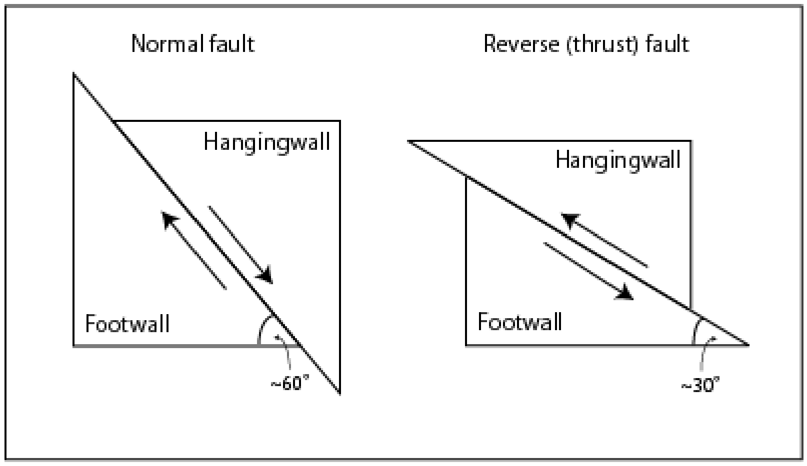
Dip-slip faults



Students look at a section of the exposed Wasatch Fault (Normal fault), Utah

Dip-slip faults can occur either as "reverse" or as "normal" faults. A normal fault occurs when the crust is extended. Alternatively such a fault can be called an extensional fault. The hanging wall moves downward, relative to the footwall. A downthrown block between two normal faults dipping towards each other is called a graben. An upthrown block between two normal faults dipping away from each other is called a horst. Low-angle normal faults with regional tectonic significance may be designated detachment faults.

A reverse fault is the opposite of a normal fault—the hanging wall moves up relative to the footwall. Reverse faults indicate shortening of the crust. The dip of a reverse fault is relatively steep, greater than 45°.



Cross-sectional illustration of normal and reverse dip-slip faults

A thrust fault has the same sense of motion as a reverse fault, but with the dip of the fault plane at less than 45°. Thrust faults typically form ramps, flats and fault-bend (hanging wall and foot wall) folds. Thrust faults form nappes and klippen in the large thrust belts.

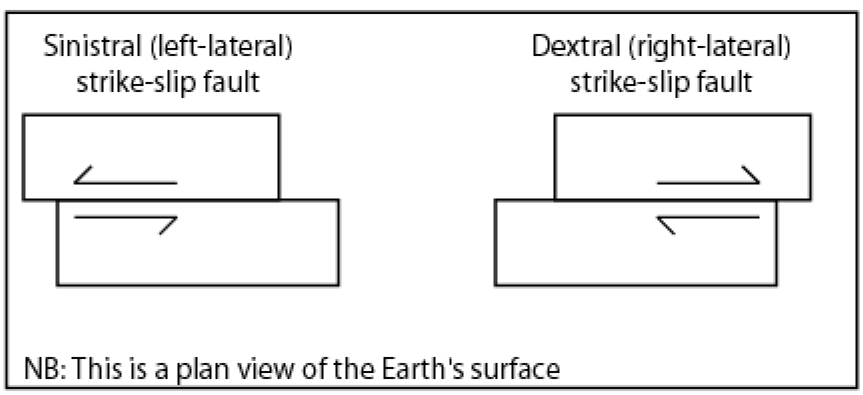
The fault plane is the plane that represents the fracture surface of a fault. Flat segments of thrust fault planes are known as *flats*, and inclined sections of the thrust are known as *ramps*. Typically, thrust faults move *within* formations by forming flats, and climb up section with ramps.

Fault-bend folds are formed by movement of the hanging wall over a non-planar fault surface and are found associated with both extensional and thrust faults.

Faults may be reactivated at a later time with the movement in the opposite direction to the original movement (fault inversion). A normal fault may therefore become a reverse fault and vice versa.

Strike-slip faults

The San Andreas Fault, a right-lateral strike-slip fault caused the massive 1906 San Francisco earthquake

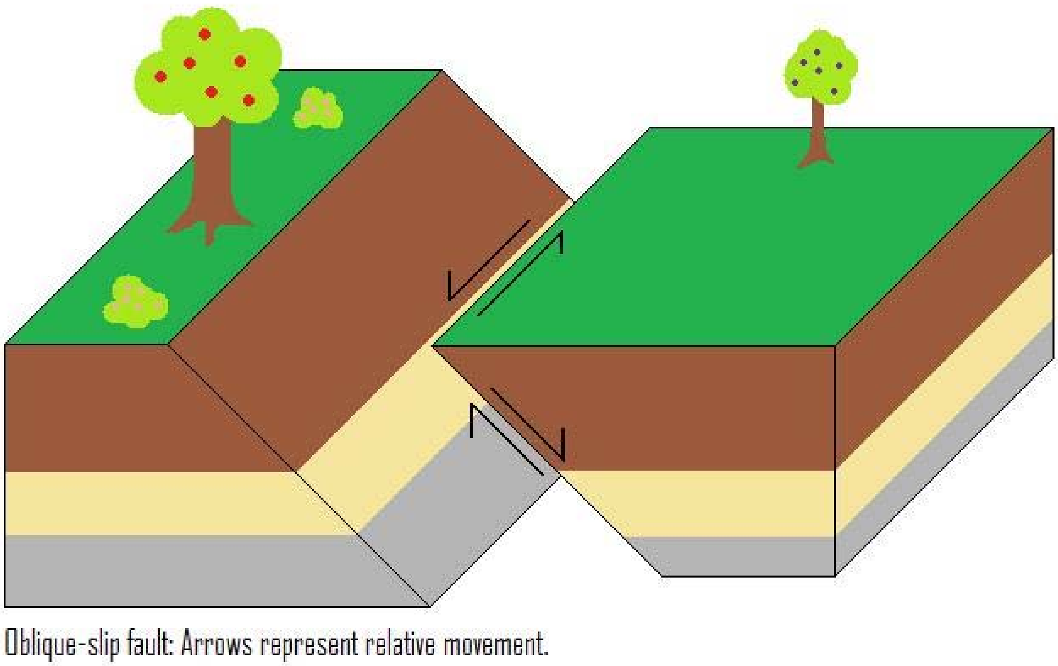


Schematic illustration of the two strike-slip fault types

The fault surface is usually near vertical and the footwall moves either left or right or laterally with very little vertical motion. **Strike-slip faults** with left-lateral motion are also known as *sinistral* faults. Those with right-lateral motion are also known as *dextral* faults.

A special class of strike-slip faults is the transform fault, where such faults form a plate boundary. These are found related to offets in spreading centers, such as mid-ocean ridges, and less commonly within continental lithosphere, such as the Alpine Fault, New Zealand. Transform faults are also referred to as conservative plate boundaries, as lithosphere is neither created or destroyed.

Oblique-slip faults



Oblique-slip fault

A fault which has a component of dip-slip and a component of strike-slip is termed an **oblique-slip fault**. Nearly all faults will have some component of both dip-slip and strike-slip, so defining a fault as oblique requires both dip and strike components to be measurable and significant. Some oblique faults occur within transtensional and transpressional regimes, others occur where the direction of extension or shortening changes during the deformation but the earlier formed faults remain active.

The *hade* angle is defined as the complement of the dip angle; it is the angle between the fault plane and a vertical plane that strikes parallel to the fault.

Listric fault

A listric fault is a type of normal fault in which fault plane is curved. The dip of the fault plane becomes shallower with increased depth.

Ring fault

Ring faults are faults that occur within collapsed volcanic calderas. Ring faults may be filled by ring dikes.

Fault rock

Salmon-colored fault gouge and associated fault separates two different rock types on the left (dark grey) and right (light grey).

Inactive fault from Sudbury to Sault Ste. Marie, Northern Ontario, Canada

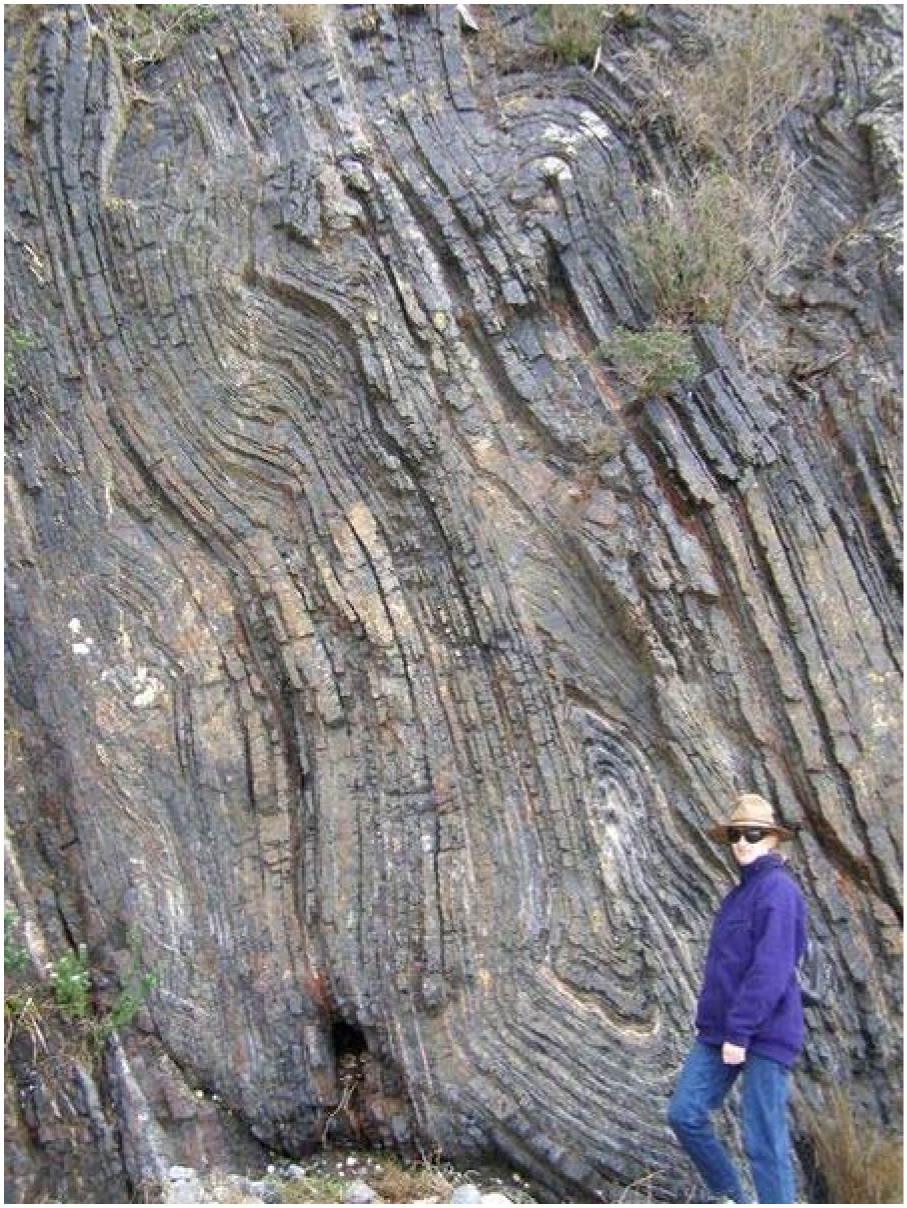
All faults have a measurable thickness, made up of deformed rock characteristic of the level in the crust where the faulting happened, of the rock types affected by the fault and of the presence and nature of any mineralising fluids. Fault rocks are classified by their textures and the implied mechanism of deformation. A fault that passes through different levels of the lithosphere will have many different types of fault rock developed along its surface. Continued dip-slip displacement tends to juxtapose fault rocks characteristic of different crustal levels, with varying degrees of overprinting. This effect is particularly clear in the case of detachment faults and major thrust faults.

The main types of fault rock include:

- Cataclasite - a fault rock which is cohesive with a poorly developed or absent planar fabric, or which is incohesive, characterised by generally angular clasts and rock fragments in a finer-grained matrix of similar composition.
    - Tectonic or Fault breccia - a medium- to coarse-grained cataclasite containing >30% visible fragments.
    - Fault gouge - an incohesive, clay-rich fine- to ultrafine-grained cataclasite, which may possess a planar fabric and containing <30% visible fragments. Rock clasts may be present
        - Clay smear - clay-rich fault gouge formed in sedimentary sequences containing clay-rich layers which are strongly deformed and sheared into the fault gouge.
- Mylonite - a fault rock which is cohesive and characterized by a well developed planar fabric resulting from tectonic reduction of grain size, and commonly containing rounded porphyroclasts and rock fragments of similar composition to minerals in the matrix
- Pseudotachylite - ultrafine-grained vitreous-looking material, usually black and flinty in appearance, occurring as thin planar veins, injection veins or as a matrix to pseudoconglomerates or breccias, which infills dilation fractures in the host rock.

Fold

A fold in Slichowice nature reserve in Kielce (Variscan orogeny)

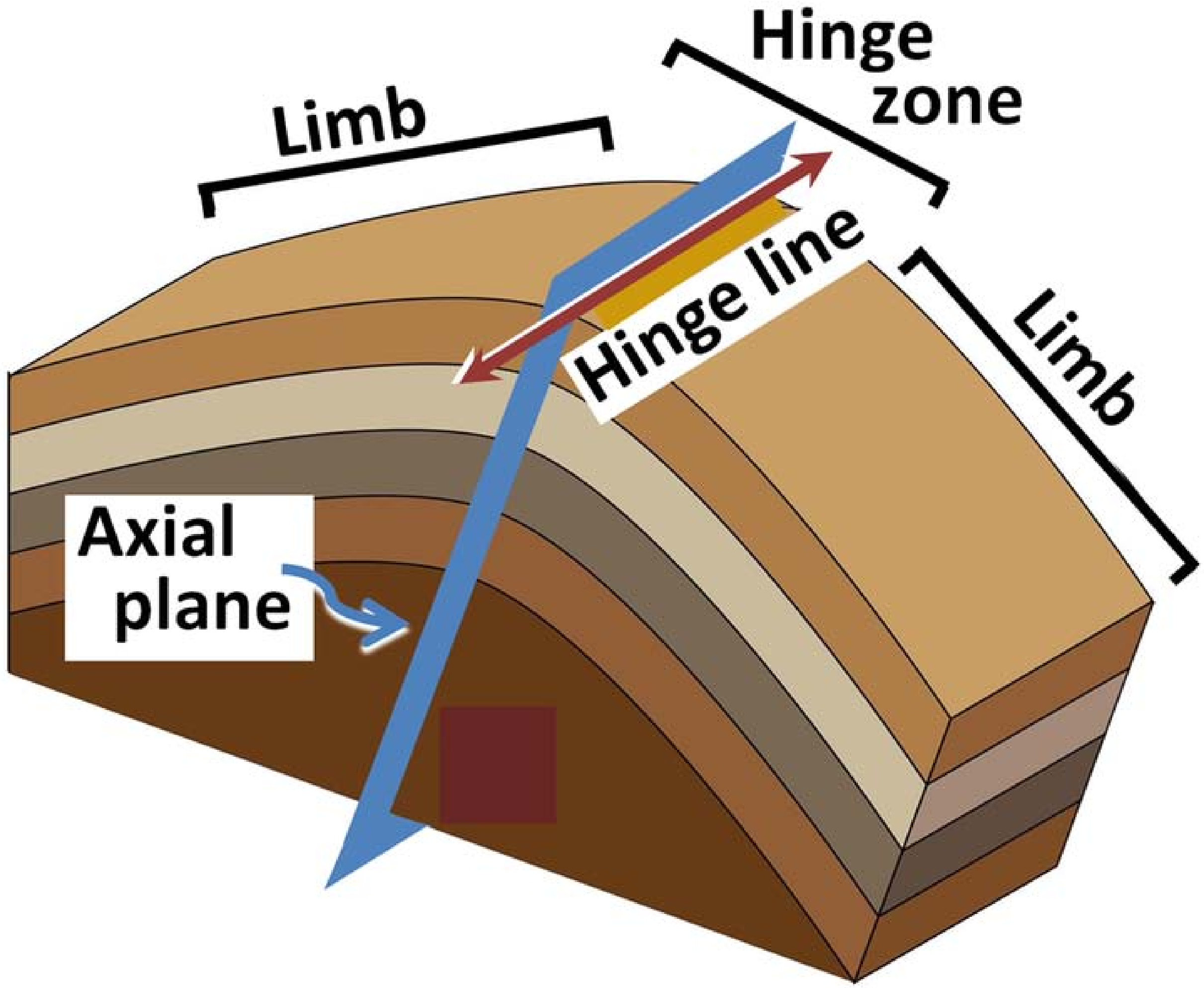


Very tight folds. Formation near Moruya, New South Wales, Australia

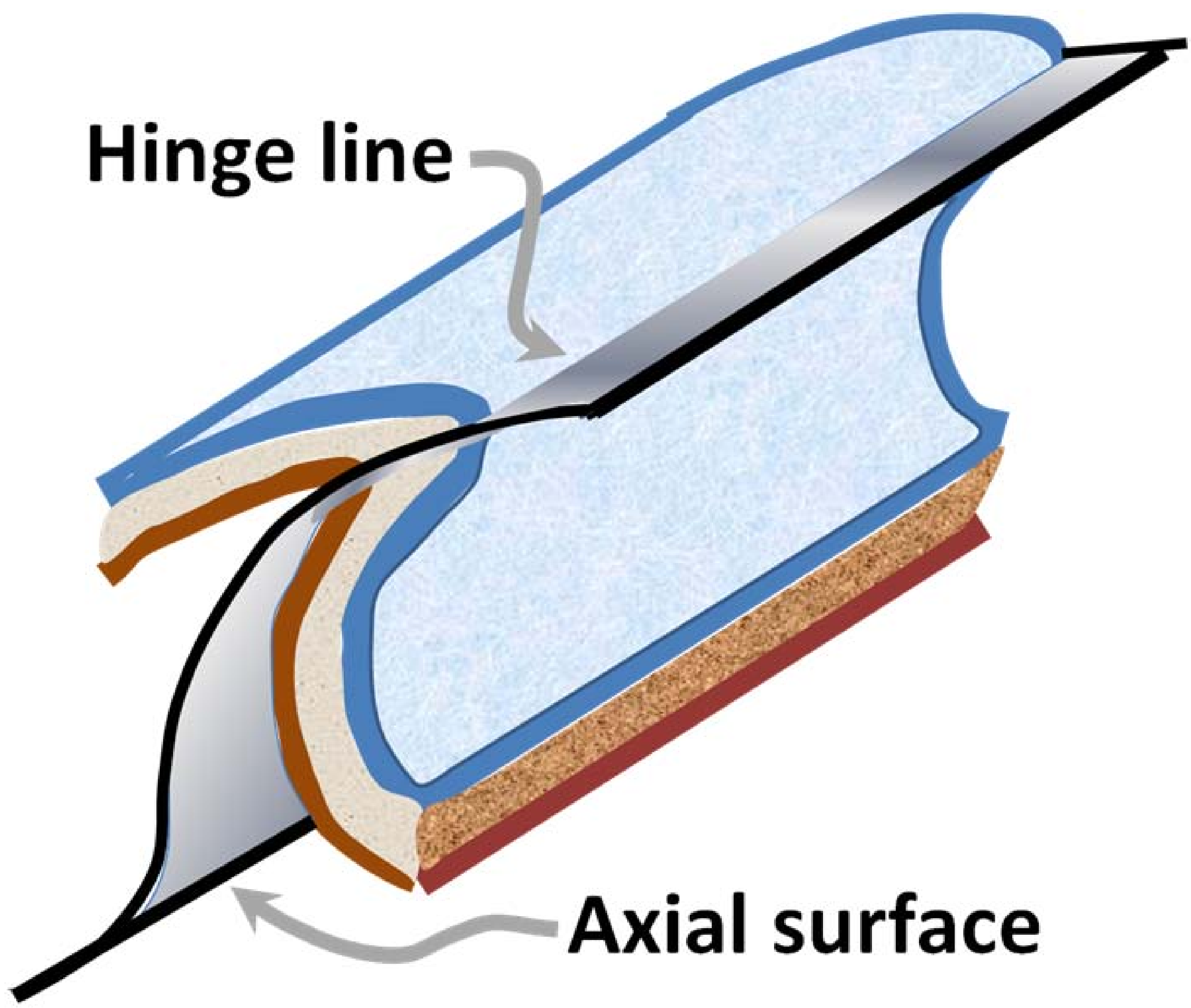
Rainbow Basin Syncline in the Barstow Formation near Barstow, California

The term **fold** is used in geology when one or a stack of originally flat and planar surfaces, such as sedimentary strata, are bent or curved as a result of plastic (that is, permanent) deformation. Synsedimentary folds are those due to slumping of sedimentary material before it is lithified. Folds in rocks vary in size from microscopic crinkles to mountain-sized folds. They occur singly as isolated folds and in extensive fold trains of different sizes, on a variety of scales. Folds form under varied conditions of stress, hydrostatic pressure, pore pressure, and temperature - hydrothermal gradient, as evidenced by their presence in soft sediments, the full spectrum of metamorphic rocks, and even as primary flow structures in some igneous rocks. A set of folds distributed on a regional scale constitutes a fold belt, a common feature of orogenic zones. Folds are commonly formed by shortening of existing layers, but may also be formed as a result of displacement on a non-planar fault (*fault bend fold*), at the tip of a propagating fault (*fault propagation fold*), by differential compaction or due to the effects of a high-level igneous intrusion e.g. above a laccolith.

Describing folds



Fold terminology. For more general fold shapes, a hinge *curve* replaces the hinge line, and a non-planar axial *surface* replaces the axial plane.



Cylindrical fold with axial surface not a plane

Folds are classified by their size, fold shape, tightness, dip of the axial plane.

Fold terminology in two dimensions

Looking at a fold surface in profile the fold can be divided into *hinge* and *limb* portions. The limbs are the flanks of the fold and the hinge is where the flanks join together. The hinge point is the point of minimum radius of curvature for a fold. The crest of the fold is the highest point of the fold surface, and the trough is the lowest point. The inflection point of a fold is the point on a limb at which the concavity reverses, on regular folds this is the mid-point of the limb.

Fold terminology in three dimensions

The hinge points along an entire folded surface form a hinge line, which can be either a *crest line* or a *trough line*. The trend and plunge of a linear hinge line gives you information about the orientation of the fold. To more completely describe the orientation of a fold, one must describe the *axial surface*. The axial surface is the surface defined by connecting all the hinge lines of stacked folding surfaces. If the axial surface is a planar surface then it is called the axial plane and can be described by the strike and dip of the plane. An *axial trace* is the line of intersection of the axial surface with any other surface (ground, side of mountain, geological cross-section).

Finally, folds can have, but don't necessarily have a *fold axis*. A fold axis, "is the closest approximation to a straight line that when moved parallel to itself, generates the form of the fold." (Davis and Reynolds, 1996 after Donath and Parker, 1964; Ramsay 1967). A fold that can be generated by a fold axis is called a *cylindrical fold*. This term has been broadened to include near-cylindrical folds. Often, the fold axis is the same as the hinge line.

Fold shape

It is necessary to convey a sense of the shape of the fold. A fold can be shaped as a chevron, with planar limbs meeting at an angular axis, as **cuspate** with curved limbs, as circular with a curved axis, or as elliptical with unequal wavelength.

Fold tightness

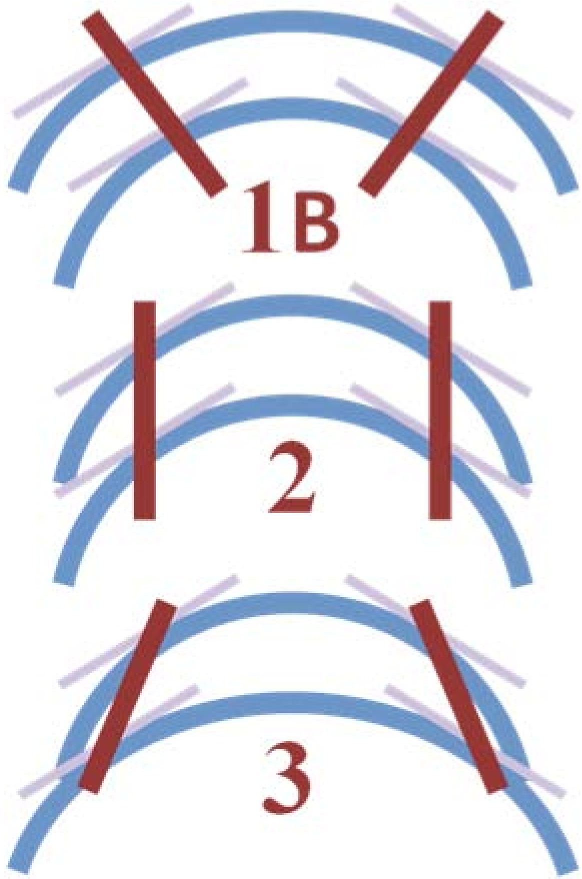
Fold tightness is defined by the angle between the fold's limbs, called the interlimb angle. Gentle folds have an interlimb angle of between 180° and 120°, open folds range from 120° to 70°, Close folds from 70° to 30°, tight folds from 30° to 0° and isoclinal folds have an interlimb angle of between 10° and zero, with essentially parallel limbs.

Fold symmetry

Not all folds are equal on both sides of the axis of the fold. Those with limbs of relatively equal length are termed symmetrical, and those with highly unequal limbs are asymmetrical. Asymmetrical folds generally have an axis at an angle to the original unfolded surface they formed on.

Deformation style classes

Folds that maintain uniform layer thickness are classed as *concentric* folds. Those that do not are called *similar folds*. Similar folds tend to display thinning of the limbs and thickening of the hinge zone. Concentric folds are caused by warping from active buckling of the layers, whereas similar folds usually form by some form of shear flow where the layers are not mechanically active. Ramsay has proposed a classification scheme for folds that often is used to describe folds in profile based upon curvature of the inner and outer lines of a fold, and the behavior of *dip isogons*. that is, lines connecting points of equal dip:



Ramsay classification of folds by convergence of dip isogons (red lines)

Ramsay classification scheme for folds

Class	Curvature *C*	Comment
1	$C_{inner} > C_{outer}$	Dip isogons converge
1A		Orthogonal thickness at hinge narrower than at limbs
1B		Parallel folds
1C		Orthogonal thickness at limbs narrower than at hinge
2	$C_{inner} = C_{outer}$	Dip isogons are parallel: **similar** folds
3	$C_{inner} < C_{outer}$	Dip isogons diverge

Fold types

Anticline – USGS

Monocline at Colorado National Monument

Recumbent fold, King Oscar Fjord

- Anticline: linear, strata normally dip away from axial center, *oldest* strata in center.
- Syncline: linear, strata normally dip toward axial center, *youngest* strata in center.
- Antiform: linear, strata dip away from axial center, age unknown, or inverted.
- Synform: linear, strata dip toward axial centre, age unknown, or inverted.
- Dome: nonlinear, strata dip away from center in all directions, *oldest* strata in center.
- Basin: nonlinear, strata dip toward center in all directions, *youngest* strata in center.
- Monocline: linear, strata dip in one direction between horizontal layers on each side.
- Chevron: angular fold with straight limbs and small hinges
- Recumbent: linear, fold axial plane oriented at low angle resulting in overturned strata in one limb of the fold.
- Slump: typically monoclinal, result of differential compaction or dissolution during sedimentation and lithification.
- Ptygmatic: Folds are chaotic, random and disconnected. Typical of sedimentary slump folding, migmatites and decollement detachment zones.
- Parasitic: short wavelength folds formed within a larger wavelength fold structure - normally associated with differences in bed thickness
- Disharmonic: Folds in adjacent layers with different wavelengths and shapes

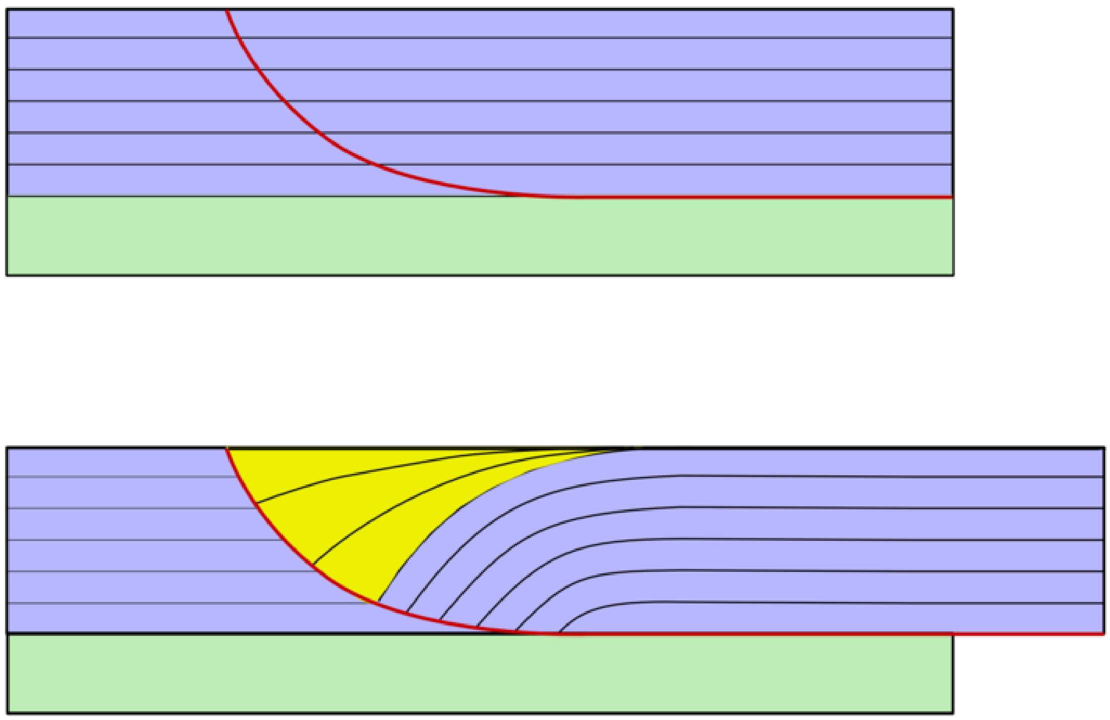
(A homocline involves strata dipping in the same direction, though not necessarily any folding.)

Causes of folding

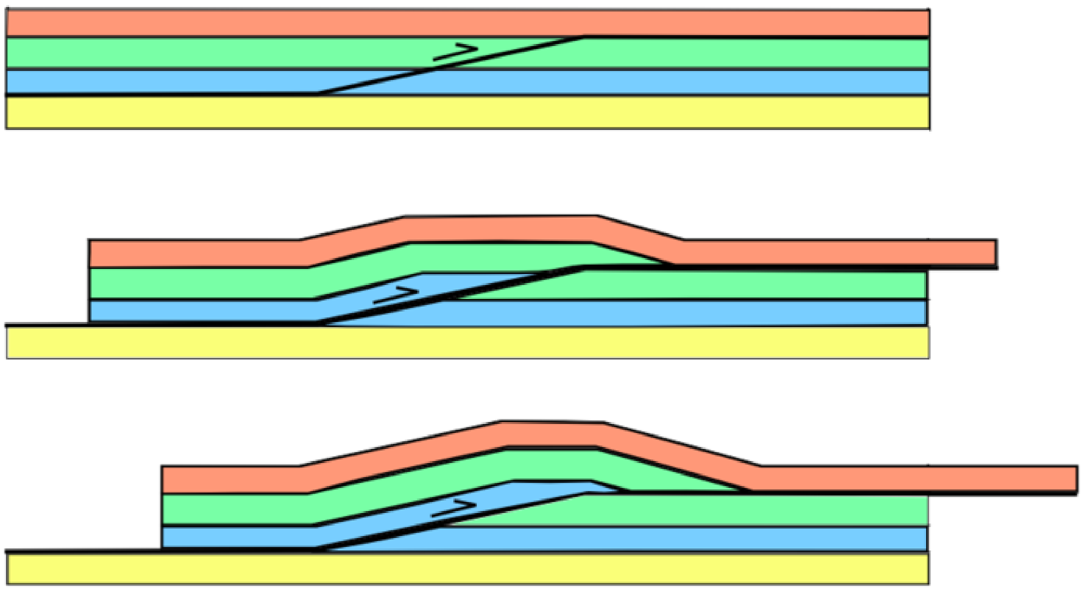
Folds appear on all scales, in all rock types, at all levels in the crust and arise from a variety of causes.

Layer-parallel shortening

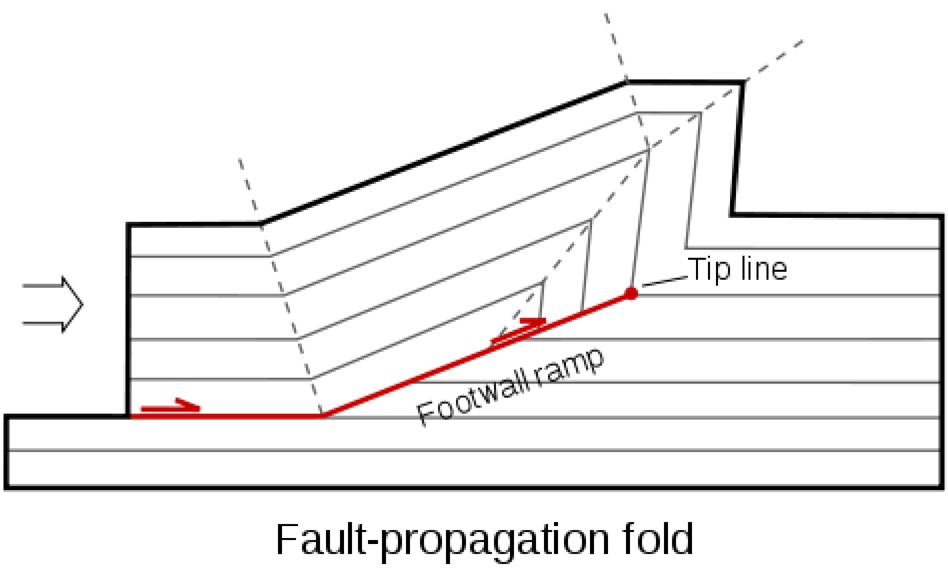
When a sequence of layered rocks is shortened parallel to its layering, this deformation may be accommodated in a number of ways, homogeneous shortening, reverse faulting or folding. The response depends on the thickness of the mechanical layering and the contrast in properties between the layers. If the layering does begin to fold, the fold style is also dependent on these properties. Isolated thick competent layers in a less competent matrix control the folding and typically generate classic rounded buckle folds accommodated by deformation in the matrix. In the case of regular alternations of layers of contrasting properties, such as sandstone-shale sequences, kink-bands, box-folds and chevron folds are normally produced.



Rollover anticline



Ramp anticline



Fault-propagation fold

Fault-related folding

Many folds are directly related to faults, associate with their propagation, displacement and the accommodation of strains between neighbouring faults.

Fault bend folding

Fault bend folds are caused by displacement along a non-planar fault. In non-vertical faults, the hanging-wall deforms to accommodate the mismatch across the fault as displacement progresses. Fault bend folds occur in both extensional and thrust faulting. In extension, listric faults form rollover anticlines in their hanging walls. In thrusting, *ramp anticlines* are formed whenever a thrust fault cuts up section from one detachment level to another. Displacement over this higher-angle ramp generates the folding.

Fault propagation folding

Fault propagation folds or *tip-line folds* are caused when displacement occurs on an existing fault without further propagation. In both reverse and normal faults this leads to folding of the overlying sequence, often in the form of a monocline.

Detachment folding

When a thrust fault continues to displace above a planar detachment without further fault propagation, detachment folds may form, typically of box-fold style. These generally occur above a good detachment such as in the Jura mountains, where the detachment occurs on middle Triassic evaporites.

Compaction

Folds can be generated in a younger sequence by differential compaction over older structures such as fault blocks and reefs.

Folding in shear zones

Shear zones that approximate to simple shear typically contain minor asymmetric folds, with the direction of overturning consistent with the overall shear sense. Some of these folds have highly curved hinge lines and are referred to as *sheath folds*. Folds in shear zones can be inherited, formed due to the orientation of pre-shearing layering or formed due to instability within the shear flow.

Sedimentary folding

Recently deposited sediments are normally mechanically weak and prone to remobilisation.

Slump folding

When slumps form in poorly consolidated sediments they commonly undergo folding, particularly at their leading edges, during their emplacement. The asymmetry of the slump folds can be used to determine paleoslope directions in sequences of sedimentary rocks.

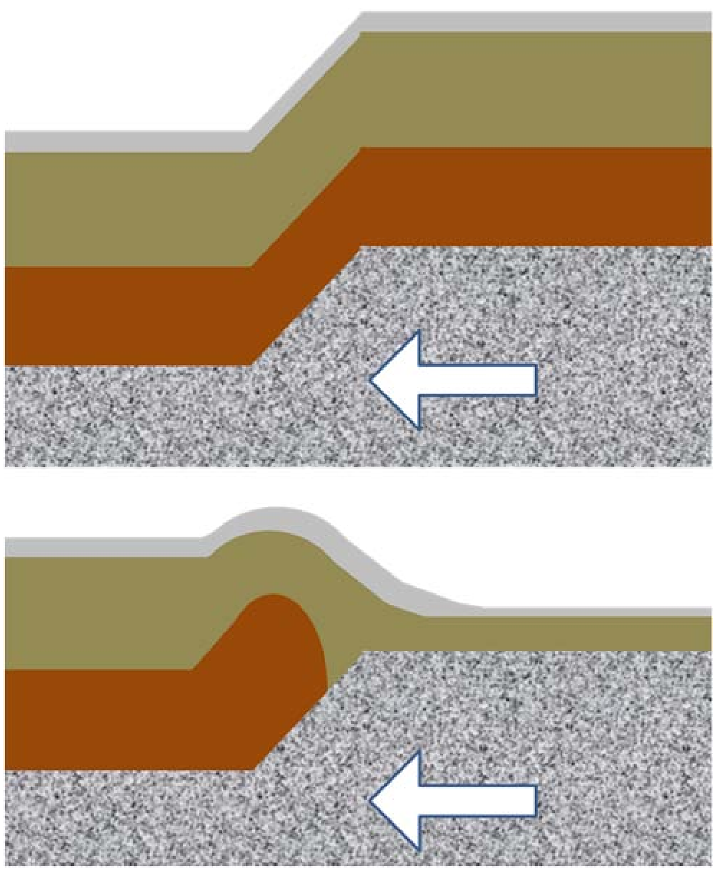
Dewatering

Rapid dewatering of sandy sediments, possibly triggered by seismic activity can cause convolute bedding.

Igneous intrusion

The emplacement of igneous intrusions tends to deform the surrounding country rock. In the case of high-level intrusions, near the Earth's surface, this deformation is concentrated above the intrusion and often takes the form of folding, as with the upper surface of a laccolith.

Flow folding



Flow folding: this picture uses artistic license to show the effect of an advancing ramp of rigid rock into compliant layers. Top: low drag by ramp: layers are not altered in thickness; Bottom: high drag: lowest layers tend to crumple.

The compliance of rock layers is referred to as *competence*: a competent layer or bed of rock can withstand an applied load without collapsing and is relatively strong, while an incompetent layer is relatively weak. When rock behaves as a fluid, as in the case of very weak rock such as rock salt, or any rock that is buried deeply enough, they typically show *flow folding* (also called *passive folding*, because little resistance is offered): the strata appear shifted undistorted, assuming any shape impressed upon them by surrounding more rigid rocks. The strata simply serve as markers of the folding. Such folding is also a feature of many igneous intrusions and glacier ice.

Folding mechanisms

Folding of rocks must balance the deformation of layers with the conservation of volume in a rock mass. This occurs by several mechanisms.

Example of a large-scale crenulation, Glengarry Basin, W.A., an example of chevron-type flexural-slip folds.

Flexural slip

Flexural slip allows folding by creating layer-parallel slip between the layers of the folded strata, which, altogether, result in deformation. The best analogy is bending a phone book, where volume preservation is accommodated by slip between the pages of the book.

Buckling

Typically, folding is thought to occur by simple buckling of a planar surface and its confining volume. The volume change is accommodated by *layer parallel shortening* the volume, which grows in *thickness*. Folding under this mechanism is typically of the similar fold style, as thinned limbs are shortened horizontally and thickened hinges do so vertically.

Mass displacement

If the folding deformation cannot be accommodated by flexural slip or volume-change shortening (buckling), the rocks are generally removed from the path of the stress. This is achieved by pressure dissolution, a form of metamorphic process, in which rocks shorten by dissolving constituents in areas of high strain and redepositing them in areas of lower strain. Folds created in this way include examples in migmatites, and areas with a strong axial planar cleavage.

Mechanics of Folding

Folds in rock are formed in relation to the stress field in which the rocks are located and the rheology, or method of response to stress, of the rock at the time at which the stress is applied.

Rheology

Weathered marble anticline at General Carrera Lake, Chile

The rheology of the layers being folded determines characteristic features of the folds that are measured in the field. Rocks that deform more easily form many short-wavelength, high-amplitude folds. Rocks that do not deform as easily form long-wavelength, low-amplitude folds.

Relationship between folds and applied stress field

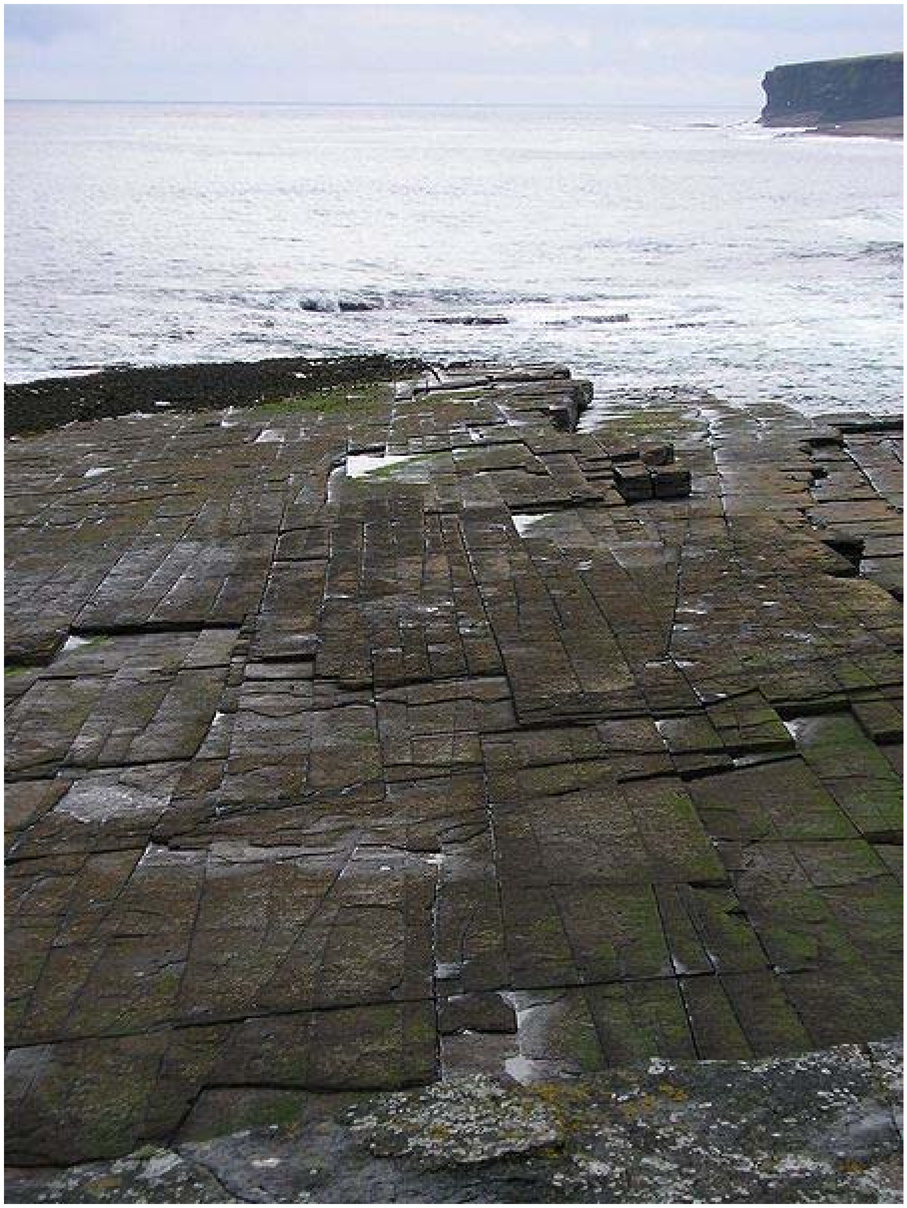
Understanding the relationship between the stress regime in which a fold forms and what structures one would expect is important in geology. Using these relationships, geologists are able to use the observed fold geometries to understand the physical forces that made them.

Chapter- 4

Joint and Shear (geology)

Joint

Columnar jointed basalt in Turkey



Joint sets on a bedding plane in flagstones, Caithness, Scotland

A rock in Abisko possibly fractured along existing joints by mechanical frost weathering

Columnar jointing in basalt, Marte Vallis, Mars

In geology the term **joint** refers to a fracture in rock where the displacement associated with the opening of the fracture is greater than the displacement due to lateral movement in the plane of the fracture (up, down or sideways) of one side relative to the other. Typically, there is little to no lateral movement across joints. This makes joints different from a fault which is defined as a fracture in rock in which one side slides laterally past the other with a displacement that is greater than the separation between the blocks on either side of the fracture. Joints normally have a regular spacing related to either the mechanical properties of the individual rock or the thickness of the layer involved. Joints generally occur as sets, with each set consisting of joints sub-parallel to each other.

Formation

Joints form in solid, hard rock that is stretched such that its brittle strength is exceeded (the point at which it breaks). When this happens the rock fractures in a plane parallel to the maximum principal stress and perpendicular to the minimum principal stress (the direction in which the rock is being stretched). This leads to the development of a single sub-parallel joint set. Continued deformation may lead to development of one or more additional joint sets. The presence of the first set strongly affects the stress orientation in the rock layer, often causing subsequent sets to form at a high angle to the first set.

Joint sets are commonly observed to have relatively constant spacing, which is roughly proportional to the thickness of the layer.

Types of joints

Joints are classified by the processes responsible for their formation, if known. ==Types with repect to formation

Tectonic joints

Tectonic joints are formed during deformation episodes whenever the differential stress is high enough to induce tensile failure of the rock, irrespective of the tectonic regime. They will often form at the same time as faults. Measurement of tectonic joint patterns can be useful in analyzing the tectonic history of an area because they give information on stress orientations at the time of formation

Unloading joints

Joints are most commonly formed when uplift and erosion removes the overlying rocks thereby reducing the compressive load and allowing the rock to expand laterally. Joints related to uplift and erosional unloading have orientations reflecting the principal stresses during the uplift. Care needs to be taken when attempting to understand past tectonic stresses to discriminate, if possible, between tectonic and unloading joints.

Exfoliation joints are special cases of unloading joints formed at, and parallel to, the current land surface in rocks of high compressive strength.

Cooling joints

Joints can also form via cooling of hot rock masses, particularly lava, forming *cooling joints*, most commonly expressed as vertical *columnar jointing*. The joint systems associated with cooling typically are polygonal because the cooling introducing stresses that are isotropic in the plane of the layer.

Types with respect to attitude and geometry:

Joints can be classified into three groups

Strike joints

Joints which run parallel to the direction of strike of country rocks are called "strike joints"

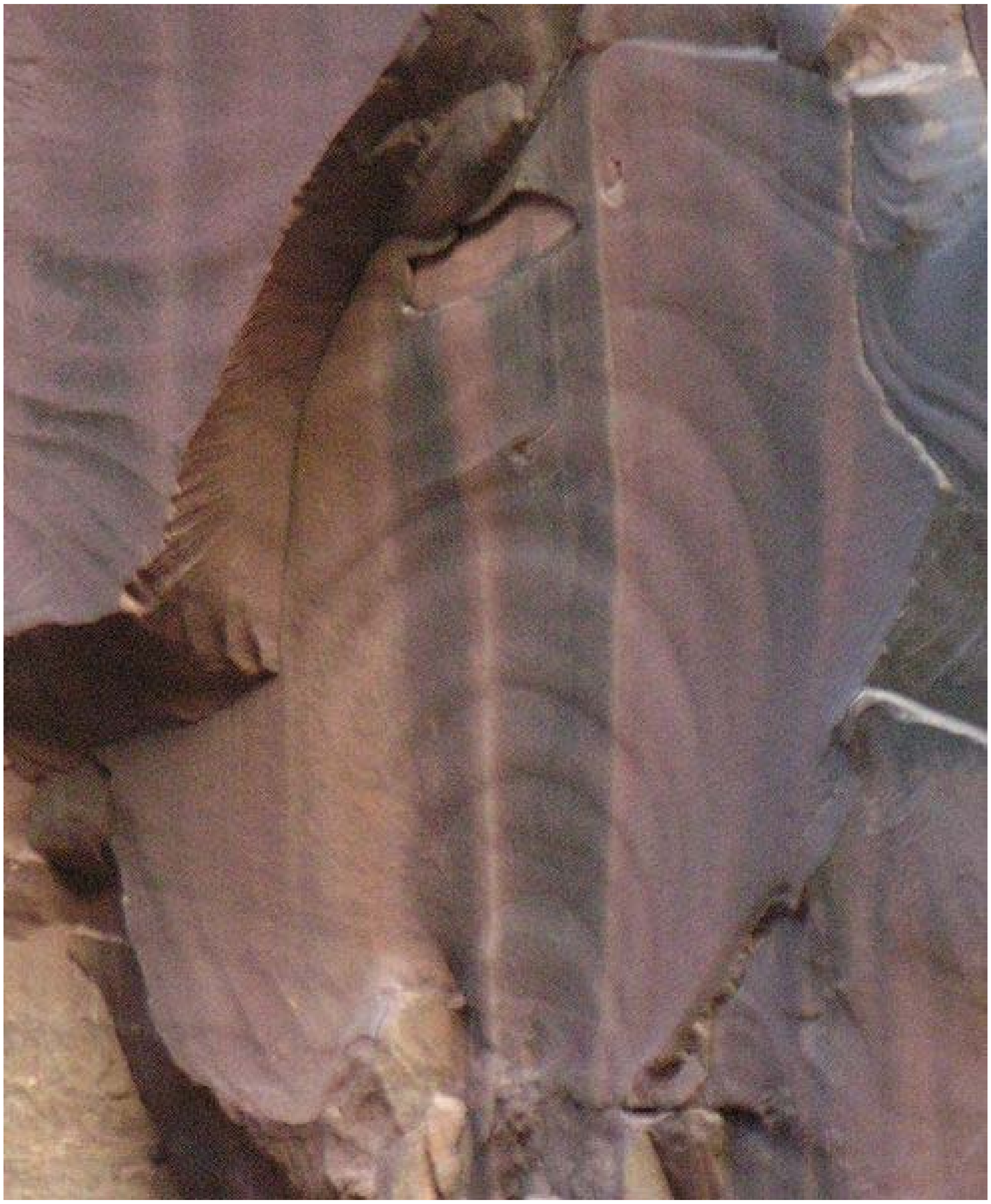
Dip joints

Joints which run parallel to the direction of dip of country rocks are called "dip joints"

Oblique joints

Joints which run oblique to the dip and strike directions of the country rocks are called "oblique joints".

Fractography



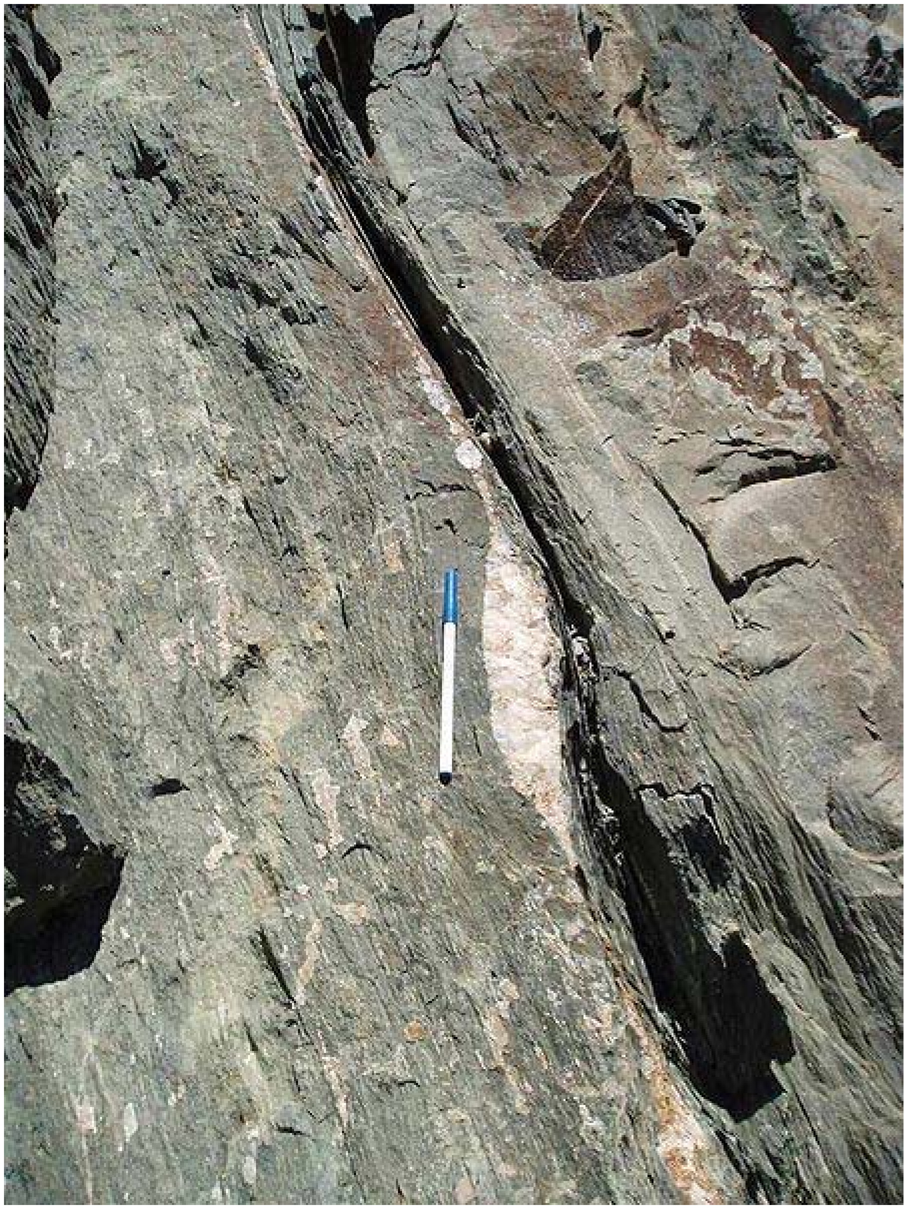
Plumose structure on a fracture surface in sandstone

Joint propagation can be studied using the techniques of fractography in which characteristic marks such as hackles and plumose structures can be used to determine propagation directions and, in some cases, the principal stress orientations.

Importance to rock strength and slope stability

Joints form one of the most important types of discontinuity within rock masses, typically having no tensile strength.

Shear



Boudinaged quartz vein(with strain fringe) showing **sinistral shear sense**, Starlight Pit, Fortnum Gold Mine, Western Australia

Shear is the response of a rock to deformation usually by compressive stress and forms particular textures. Shear can be homogeneous or non-homogeneous, and may be pure shear or simple shear. Study of geological shear is related to the study of structural geology, rock microstructure or rock texture and fault mechanics.

The process of shearing occurs within brittle, brittle-ductile, and ductile rocks. Within purely brittle rocks, compressive stress results in fracturing and simple faulting.

Rocks

Rocks typical of shear zones include mylonite, cataclasite, S-tectonite and L-tectonite, pseudotachylite, certain breccias and highly foliated versions of the wall rocks.

Shear zone

Asymmetric shear in basalt, Labouchere mine, Glengarry Basin, WA. Shear asymmetry is dextral, pen for scale

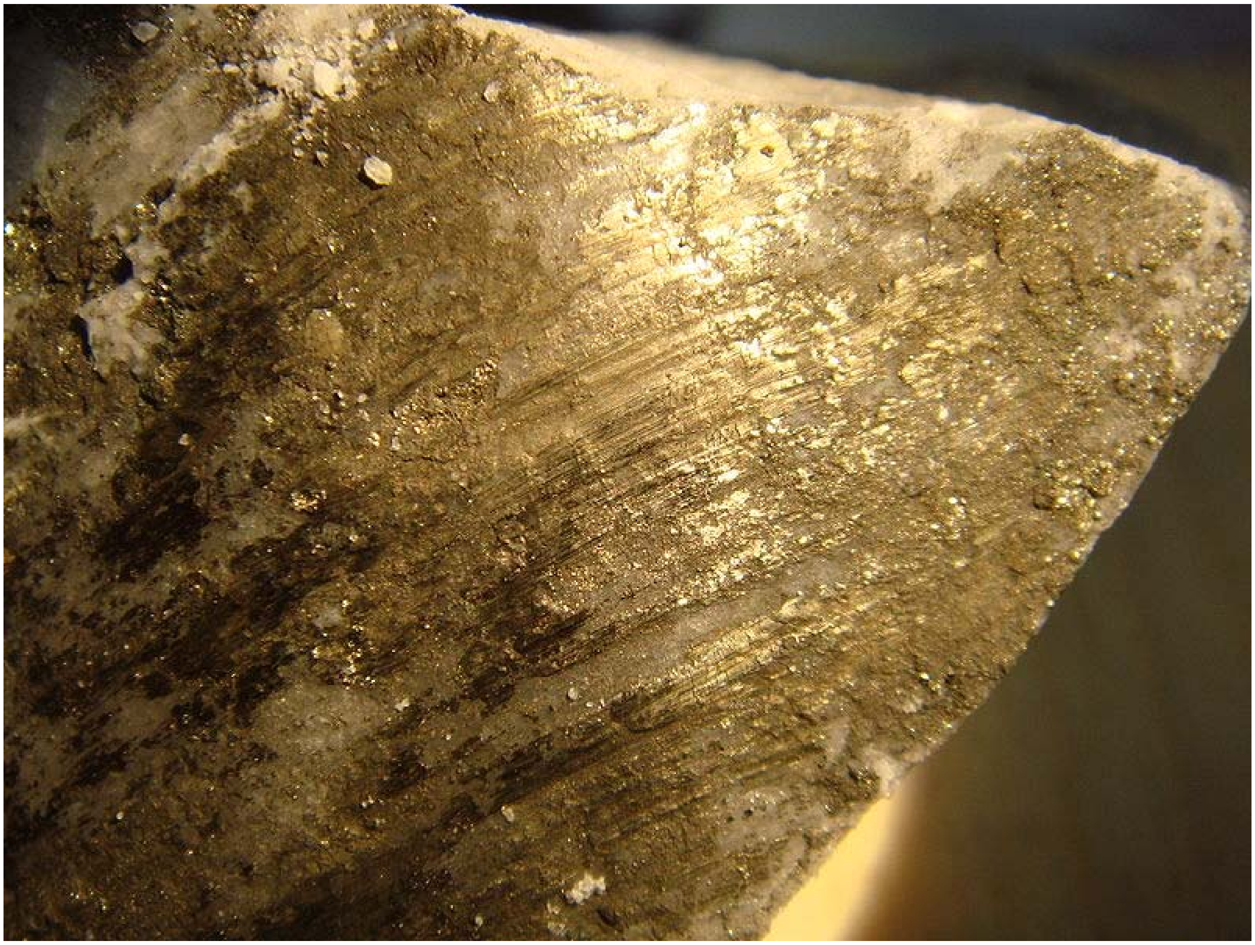
A shear zone is a tabular to sheetlike, planar or curviplanar zone composed of rocks that are more highly strained than rocks adjacent to the zone. Typically this is a type of fault but it may be difficult to place a distinct fault plane into the shear zone. Shear zones may form zones of much more intense foliation, deformation, and folding. En echelon veins or fractures may be observed within shear zones.

Many shear zones host ore deposits as they are a focus for hydrothermal flow through orogenic belts. They may often show some form of retrograde metamorphism from a peak metamorphic assemblage and are commonly metasomatised.

Shear zones can be only inches wide, or up to several kilometres wide. Often, due to their structural control and presence at the edges of tectonic blocks, shear zones are mappable units and form important discontinuities to separate terranes. As such, many large and long shear zones are named, identical to fault systems.

When the horizontal displacement of this faulting can be measured in the tens or hundreds of kilometers of length, the fault is referred to as a megashear. Megashears often indicate the edges of ancient tectonic plates.

Mechanisms of shearing



Dextral slickenside of pyrite

The mechanisms of shearing depend on the pressure and temperature of the rock and on the rate of shear which the rock is subjected to. The response of the rock to these conditions determines how it accommodates the deformation.

Shear zones which occur in more brittle rheological conditions (cooler, less confining pressure) or at high rates of strain, tend to fail by brittle failure; breaking of minerals, which are ground up into a breccia with a *milled* texture.

Shear zones which occur under brittle-ductile conditions can accommodate much deformation by enacting a series of mechanisms which rely less on fracture of the rock and occur within the minerals and the mineral lattices themselves. Shear zones accommodate compressive stress by movement on foliation planes.

Shearing at ductile conditions may occur by and *dislocation creep* within minerals, by fracturing of minerals and growth of sub-grain boundaries, as well as by *lattice glide*, particularly on platy minerals, especially micas.

Mylonites are essentially ductile shear zones.

Microstructures of shear zones

Typical example of dextral shear foliation in an L-S tectonite, with pencil pointing in direction of shear sense. Note the sinusoidal nature of the shear foliation.

During the initiation of shearing, a penetrative planar foliation is first formed within the rock mass. This manifests as realignment of textural features, growth and realignment of micas and growth of new minerals.

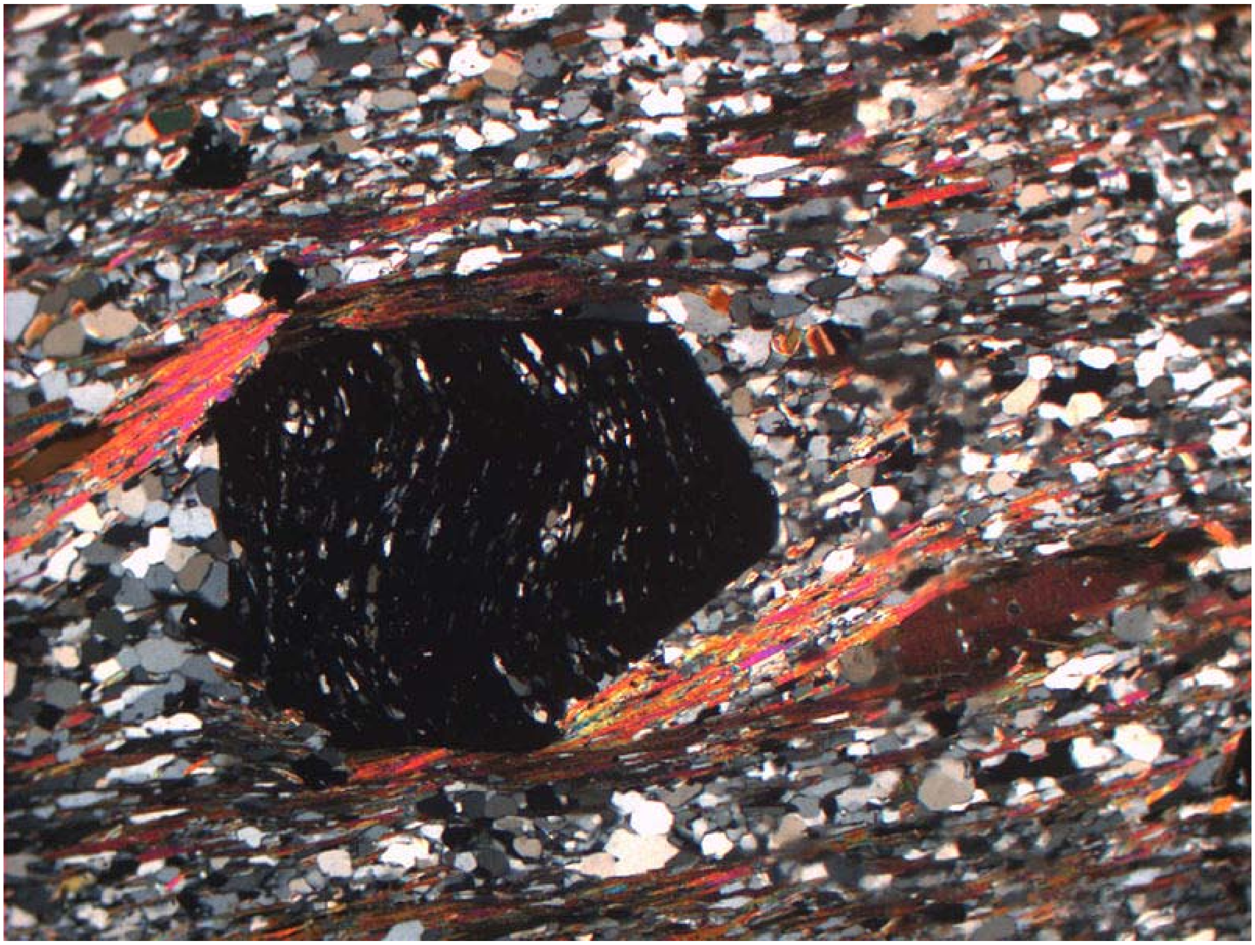
The incipient shear foliation typically forms normal to the direction of principal shortening, and is diagnostic of the direction of shortening. In symmetric shortening, objects flatten on this shear foliation much the same way that a round ball of treacle flattens with gravity.

Within asymmetric shear zones, the behavior of an object undergoing shortening is analogous to the ball of treacle being smeared as it flattens, generally into an ellipse. Within shear zones with pronounced displacements a shear foliation may form at a shallow angle to the gross plane of the shear zone. This foliation ideally manifests as a sinusoidal set of foliations formed at a shallow angle to the main shear foliation, and which curve into the main shear foliation. Such rocks are known as L-S tectonites.

If the rock mass begins to undergo large degrees of lateral movement, the strain ellipse lengthens into a cigar shaped volume. At this point shear foliaions begin to break down into a rodding lineation or a stretch lineation. Such rocks are known as L-tectonites.

Stretched pebble conglomerate L-tectonite illustrating a stretch lineation within a shear zone, Glengarry Basin, Australia. Pronounced asymmetric shearing has stretched the conglomerate pebbles into elongate cigar shaped rods.

Ductile shear microstructures



Thin section (crossed polars) of Garnet-Mica-Schist showing a rotated porphyroblast of garnet, mica fish and elongated minerals. This specimen was from close to a shear zone in Norway (the Ose thrust), the garnet in the centre (black) is approximately 2mm in diameter

Very distinctive textures form as a consequence of ductile shear. An important group of microstructures observed in ductile shear zones are S-planes, C-planes and C' planes.

- S-planes or *schistosité* planes are generally defined by a planar fabric caused by the alignment of micas or platy minerals. Define the flattened long-axis of the strain ellipse.
- C-planes or *cisaillement* planes form parallel to the shear zone boundary. The angle between the C and S planes is always acute, and defines the shear sense. Generally, the lower the C-S angle the greater the strain.
- The C' planes, also known as shear bands and secondary shear fabrics, are commonly observed in strongly foliated mylonites especially phyllonites, and form at an angle of about 20 degrees to the S-plane.

The sense of shear shown by both S-C and S-C' structures matches that of the shear zone in which they are found.

Other microstructures which can give sense of shear include:

- sigmoidal veins
- mica fish
- rotated porphyroclasts
- asymmetric boudins (Figure 1)
- asymmetric folds

Transpression

Transpression regimes are formed during oblique collision of tectonic plates and during non-orthogonal subduction. Typically a mixture of oblique-slip thrust faults and strike-slip or transform faults are formed. Microstructural evidence of transpressional regimes can be rodding lineations, mylonites, augen-structured gneisses, mica fish and so on.

A typical example of a transpression regime is the Alpine Fault zone of New Zealand, where the oblique subduction of the Pacific Plate under the Indo-Australian Plate is converted to oblique strike-slip movement. Here, the orogenic belt attains a trapezoidal shape dominated by oblique splay faults, steeply-dipping recumbent nappes and fault-bend folds.

The Alpine Schist of New Zealand is characterised by heavily crenulated and sheared phyllite. It s being pushed up at the rate of 8 to 10 mm per year, and the area is prone to large earthquakes with a south block up and west oblique sense of movement.

Transtension

Transtension regimes are oblique tensional environments. Oblique, normal geologic fault and detachment faults in rift zones are the typical structural manifestations of transtension conditions. Microstructural evidence of transtension includes rodding or stretching lineations, stretched porphyroblasts, mylonites, etc.

Chapter- 5

Types of Rock and their Formation

1. Sedimentary rock

Middle Triassic marginal marine sequence of siltstones (below) and limestones (above), Virgin Formation, southwestern Utah.

Sedimentary rock is a type of rock that is formed by sedimentation of material at the Earth's surface and within bodies of water. Sedimentation is the collective name for processes that cause mineral and/or organic particles (detritus) to settle and accumulate or minerals to precipitate from a solution. Particles that form a sedimentary rock by accumulating are called sediment. Before being deposited, sediment was formed by weathering and erosion in a source area, and then transported to the place of deposition by water, wind, mass movement or glaciers which are called agents of denudation.

The sedimentary rock cover of the continents of the Earth's crust is extensive, but the total contribution of sedimentary rocks is estimated to be only 5% of the total volume of the crust. Sedimentary rocks are only a thin veneer over a crust consisting mainly of igneous and metamorphic rocks.

Sedimentary rocks are deposited in layers as strata, forming a structure called bedding. The study of sedimentary rocks and rock strata provides information about the subsurface that is useful for civil engineering, for example in the construction of roads, houses, tunnels, canals or other constructions. Sedimentary rocks are also important sources of natural resources like coal, fossil fuels, drinking water or ores.

The study of the sequence of sedimentary rock strata is the main source for scientific knowledge about the Earth's history, including palaeogeography, paleoclimatology and the history of life.

The scientific discipline that studies the properties and origin of sedimentary rocks is called sedimentology. Sedimentology is both part of geology and physical geography and overlaps partly with other disciplines in the Earth sciences, such as pedology, geomorphology, geochemistry or structural geology.

Classification

Sedimentary rocks are classified into three groups. These groups are clastic, chemical precipitate and biochemical (or biogenic).

Clastic rock

Clastic rocks are composed of fragments, or **clasts**, of pre-existing rock. Geologists use the term **clastic** with reference to sedimentary rocks as well as to particles in sediment transport whether in suspension or as bed load, and in sediment deposits.

Clastic metamorphic and igneous rocks

Clastic metamorphic rocks include breccias formed in faults, as well as some protomylonite and pseudotachylite. Occasionally, metamorphic rocks can be brecciated via hydrothermal fluids, forming a hydrofracture breccia.

Clastic igneous rocks include pyroclastic volcanic rocks such as tuff, agglomerate and intrusive breccias, as well as some marginal eutaxitic and taxitic intrusive morphologies. Igneous clastic rocks are broken by flow, injection or explosive disruption of solid or semi-solid igneous rocks or lavas.

Clastic sediments

Claystone from Montana

Clastic sedimentary rocks are rocks composed predominantly of broken pieces or *clasts* of older weathered and eroded rocks. Clastic sediments or sedimentary rocks are classified based on grain size, clast and cementing material (matrix) composition, and texture. The classification factors are often useful in determining a sample's environment of deposition.

The walls of Lower Antelope Canyon are composed of sandstone, a common sedimentary rock

Grain size determines the basic name of a clastic sedimentary rock. Grain size varies from clay in shales and claystones; through silt in siltstones; sand in sandstones; and gravel, cobble, to boulder sized fragments in conglomerates and breccias. The Krumbein phi (φ) scale numerically orders these terms in a logarithmic size scale.

Composition includes the chemical and mineralogic make-up of the single or varied fragments and the cementing material (matrix) holding the clasts together as a rock. These differences are most commonly used in the framework grains of sandstones. Sandstones rich in quartz are called quartz arenites, those rich in feldspar are called arkoses, and those rich in lithics are called lithic sandstones.

An example clastic environment would be a river system in which the full range of grains being transported by the moving water consist of pieces eroded from solid rock upstream.

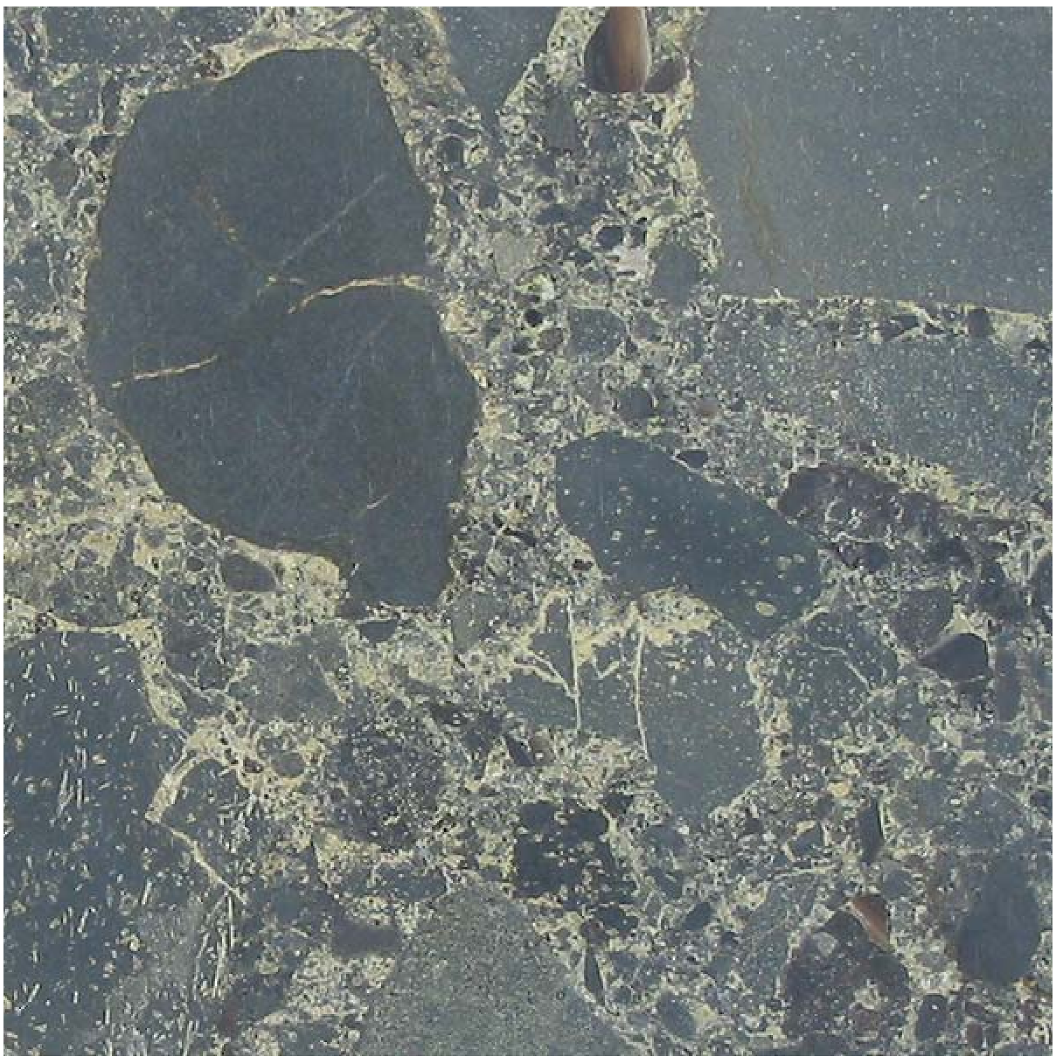
Sedimentary breccias

Sedimentary breccias are a type of clastic sedimentary rock which are composed of angular to subangular, randomly oriented clasts of other sedimentary rocks. They may form either

1. in submarine debris flows, avalanches, mud flow or mass flow in an aqueous medium. Technically, turbidites are a form of debris flow deposit and are a fine-grained peripheral deposit to a sedimentary breccia flow.
2. as angular, poorly sorted, very immature fragments of rocks in a finer grained groundmass which are produced by mass wasting. These are, in essence, lithified colluvium. Thick sequences of sedimentary (colluvial) breccias are generally formed next to fault scarps in grabens.

In the field, it may at times be difficult to distinguish between a debris flow sedimentary breccia and a colluvial breccia, especially if one is working entirely from drilling information. Sedimentary breccias are an integral host rock for many sedimentary exhalative deposits.

Igneous clastic rocks



Basalt breccia, green groundmass is composed of epidote

Igneous clastic rocks can be divided into two classes:

1. Broken, fragmental rocks produced by intrusive processes, usually associated with plutons or porphyry stocks
2. Broken, fragmental rocks associated with volcanic eruptions, both of lava and pyroclastic type

Hydrothermal clastic rocks

Hydrothermal clastic rocks are generally restricted to those formed by hydrofracture, the process by which hydrothermal circulation cracks and brecciates the wall rocks and fills it in with veins. This is particularly prominent in epithermal ore deposits and is associated with alteration zones around many intrusive rocks, especially granites. Many skarn and greisen deposits are associated with hydrothermal breccias.

Impact breccias

A fairly rare form of clastic rock may form during meteorite impact. This is composed primarily of ejecta; clasts of country rock, melted rock fragments, tektites (glass ejected from the impact crater) and exotic fragments, including fragments derived from the impactor itself.

Identifying a clastic rock as an impact breccia requires recognising shatter cones, tektites, spherulites, and the morphology of an impact crater, as well as potentially recognizing particular chemical and trace element signatures, especially osmiridium.

Organic

Outcrop of Ordovician oil shale (kukersite), northern Estonia

Organic sedimentary rocks contain materials generated by living organisms, and include carbonate minerals created by organisms, such as corals, mollusks, and foraminifera, which cover the ocean floor with layers of calcium carbonate, which can later form limestone. Other examples include stromatolites, the flint nodules found in chalk (which is itself a biochemical sedimentary rock, a form of limestone), and coal and oil shale (derived from the remains of tropical plants and subjected to heat).

Chemical

Chemical sedimentary rocks form when minerals in solution become supersaturated and precipitate. In marine environments, this is a method for the formation of limestone. Another common environment in which chemical sedimentary rocks form is a body of water that is evaporating. Evaporation decreases the amount of water without decreasing the amount of dissolved material. Therefore, the dissolved material can become oversaturated and precipitate. Sedimentary rocks from this process can include the evaporite minerals halite (rock salt), sylvite, barite and gypsum.

Formation

Cross-bedding and scour in a fine sandstone; the Logan Formation (Mississippian) of Jackson County, Ohio.

Sedimentary rocks are formed when sediment is deposited out of air, ice, wind, gravity, or water flows carrying the particles in suspension. This sediment is often formed when weathering and erosion break down a rock into loose material in a source area. The material is then transported from the source area to the deposition area. The type of sediment transported depends on the geology of the hinterland (the source area of the sediment). However, some sedimentary rocks, like evaporites, are composed of material that formed at the place of deposition. The nature of a sedimentary rock therefore not only depends on sediment supply, but also on the sedimentary depositional environment in which it formed.

Sedimentary environments

The setting in which a sedimentary rock forms is called the sedimentary environment. Every environment has a characteristic combination of geologic processes and circumstances. The type of sediment that is deposited is not only dependent on the sediment that is transported to a place, but also on the environment itself.

A marine environment means the rock was formed in a sea or ocean. Often, a distinction is made between deep and shallow marine environments. Deep marine usually refers to environments more than 200 m below the water surface. Shallow marine environments exist adjacent to coastlines and can extend out to the boundaries of the continental shelf. The water in such environments has a generally higher energy than that in deep environments, because of wave activity. This means coarser sediment particles can be transported and the deposited sediment can be coarser than in deep environments. When the available sediment is transported from the continent, an alternation of sand, clay and silt is deposited. When the continent is far away, the amount of such sediment brought in may be small, and biochemical processes dominate the type of rock that forms. Especially in warm climates, shallow marine environments far offshore mainly see deposition of carbonate rocks. The shallow, warm water is an ideal habitat for many small organisms that build carbonate skeletons. When these organisms die their skeletons sink to the bottom, forming a thick layer of calcareous mud that may lithify into limestone. Warm shallow marine environments also are ideal environments for coral reefs, where the sediment consists mainly of the calcareous skeletons of larger organisms.

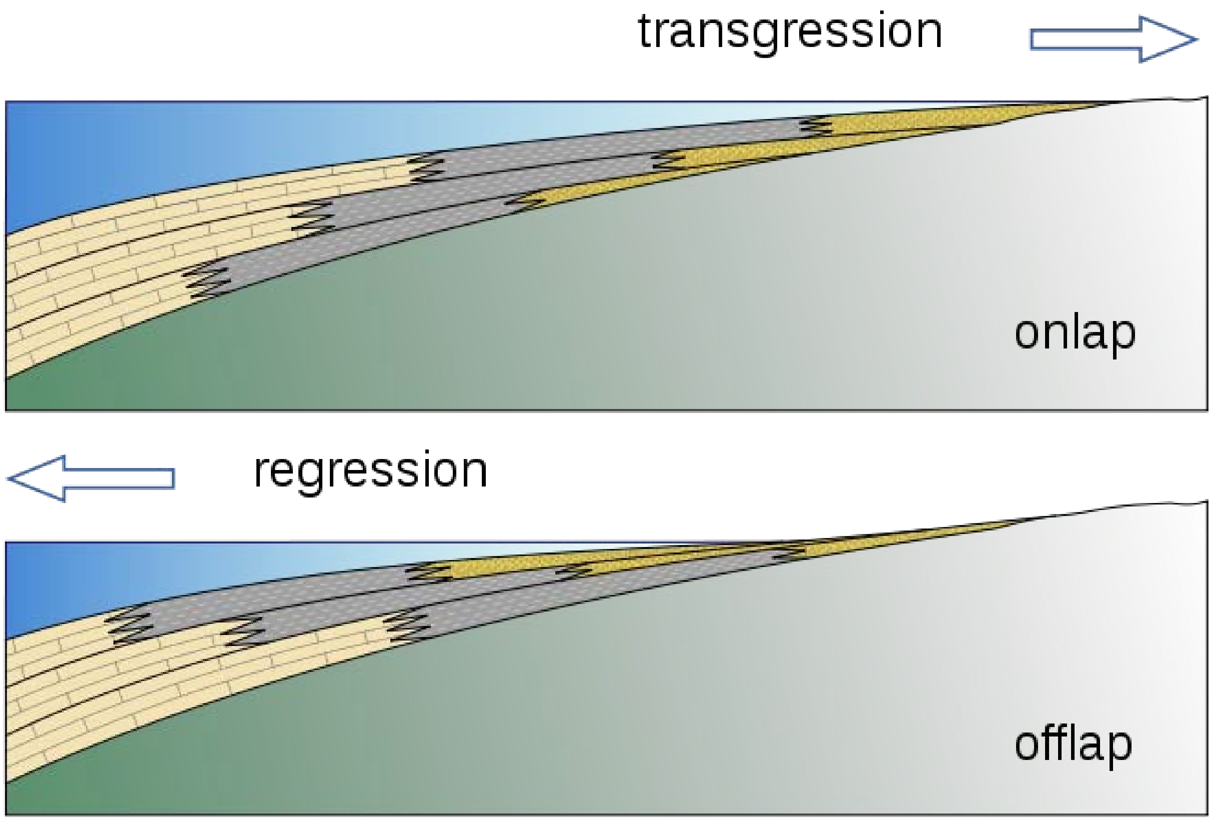
In deep marine environments, the water current over the sea bottom is small. Only fine particles can be transported to such places. Typically sediments depositing on the ocean floor are fine clay or small skeletons of micro-organisms. At 4 km depth, the solubility of carbonates increases dramatically (the depth zone where this happens is called the lysocline). Calcareous sediment that sinks below the lysocline dissolve, so no limestone can be formed below this depth. Skeletons of micro-organisms formed of silica (such as radiolarians) still deposit though. An example of a rock formed out of silica skeletons is radiolarite. When the bottom of the sea has a small inclination, for example at the continental slopes, the sedimentary cover can become unstable, causing turbidity currents. Turbidity currents are sudden disturbances of the normally quite deep marine environment and can cause the geologically speaking instantaneous deposition of large amounts of sediment, such as sand and silt. The rock sequence formed by a turbidity current is called a turbidite.

The coast is an environment dominated by wave action. At the beach, dominantly coarse sediment like sand or gravel is deposited, often mingled with shell fragments. Tidal flats and shoals are places that sometimes dry out because of the tide. They are often cross-cut by gullies, where the current is strong and the grain size of the deposited sediment is larger. Where along a coast (either the coast of a sea or a lake) rivers enter the body of water, deltas can form. These are large accumulations of sediment transported from the continent to places in front of the mouth of the river. Deltas are dominantly composed of clastic sediment.

A sedimentary rock formed on the land has a continental sedimentary environment. Examples of continental environments are lagoons, lakes, swamps, floodplains and alluvial fans. In the quiet water of swamps, lakes and lagoons, fine sediment is deposited, mingled with organic material from dead plants and animals. In rivers, the energy of the water is much higher and the transported material consists of clastic sediment. Besides transport by water, sediment can in continental environments also be transported by wind or glaciers. Sediment transported by wind is called aeolian and is always very well sorted, while sediment transported by a glacier is called glacial and is characterized by very poor sorting.

Sedimentary facies

Sedimentary environments usually exist alongside each other in certain natural successions. A beach, where sand and gravel is deposited, is usually bounded by a deeper marine environment a little offshore, where finer sediments are deposited at the same time. Behind the beach, there can be dunes (where the dominant deposition is well sorted sand) or a lagoon (where fine clay and organic material is deposited). Every sedimentary environment has its own characteristic deposits. The typical rock formed in a certain environment is called its sedimentary facies. When sedimentary strata accumulate through time, the environment can shift, forming a change in facies in the subsurface at one location. On the other hand, when a rock layer with a certain age is followed laterally, the lithology (the type of rock) and facies eventually change.



Shifting sedimentary facies in the case of transgression (above) and regression of the sea (below).

Facies can be distinguished in a number of ways: the most common ways are by the lithology (for example: limestone, siltstone or sandstone) or by fossil content. Coral for example only lives in warm and shallow marine environments and fossils of coral are thus typical for shallow marine facies. Facies determined by lithology are called lithofacies; facies determined by fossils are biofacies.

Sedimentary environments can shift their geographical positions through time. Coastlines can shift in the direction of the sea when the sea level drops, when the surface rises due to tectonic forces in the Earth's crust or when a river forms a large delta. In the subsurface, such geographic shifts of sedimentary environments of the past are recorded in shifts in sedimentary facies. This means that sedimentary facies can change either parallel or perpendicular to an imaginary layer of rock with a fixed age, a phenomenon described by Walther's facies rule.

The situation in which coastlines move in the direction of the continent is called transgression. In the case of transgression, deeper marine facies are deposited over shallower facies, a succession called onlap. Regression is the situation in which a coastline moves in the direction of the sea. With regression, shallower facies are deposited on top of deeper facies, a situation called offlap.

The facies of all rocks of a certain age can be plotted on a map to give an overview of the palaeogeography. A sequence of maps for different ages can give an insight in the development of the regional geography.

Sedimentary basins

Places where large-scale sedimentation takes place are called sedimentary basins. The amount of sediment that can be deposited in a basin depends on the depth of the basin, the so called accommodation space. Depth, shape and size of a basin depend on tectonics, movements within the Earth's lithosphere. Where the lithosphere moves upward (tectonic uplift), land eventually rises above sea level, so that and erosion removes material, and the area becomes a source for new sediment. Where the lithosphere moves downward (tectonic subsidence), a basin forms and sedimentation can take place. When the lithosphere keeps subsiding, new accommodation space keeps being created.

A type of basin formed by the moving apart of two pieces of a continent is called a rift basin. Rift basins are elongated, narrow and deep basins. Due to divergent movement, the lithosphere is stretched and thinned, so that the hot asthenosphere rises and heats the overlying rift basin. Apart from continental sediments, rift basins normally also have part of their infill consisting of volcanic deposits. When the basin grows due to continued stretching of the lithosphere, the rift grows and the sea can enter, forming marine deposits.

When a piece of lithosphere that was heated and stretched cools again, its density rises, causing isostatic subsidence. If this subsidence continues long enough the basin is called a sag basin. Examples of sag basins are the regions along passive continental margins, but sag basins can also be found in the interior of continents. In sag basins, the extra weight of the newly deposited sediments is enough to keep the subsidence going in a vicious circle. The total thickness of the sedimentary infill in a sag basins can thus exceed 10 km.

A third type of basin exists along convergent plate boundaries - places where one tectonic plate moves under another into the asthenosphere. The subducting plate bends and forms a fore-arc basin in front of the overriding plate—an elongated, deep asymmetric basin. Fore-arc basins are filled with deep marine deposits and thick sequences of turbidites. Such infill is called flysch. When the convergent movement of the two plates results in continental collision, the basin becomes shallower and develops into a foreland basin. At the same time, tectonic uplift forms a mountain belt in the overriding plate, from which large amounts of material are eroded and transported to the basin. Such erosional material of a growing mountain chain is called molasse and has either a shallow marine or a continental facies.

At the same time, the growing weight of the mountain belt can cause isostatic subsidence in the area of the overriding plate on the other side to the mountain belt. The basin type resulting from this subsidence is called a back-arc basin and is usually filled by shallow marine deposits and molasse.

Cyclic alternation of competent and less competent beds in the Blue Lias at Lyme Regis, southern England.

Influence of astronomical cycles

In many cases facies changes and other lithological features in sequences of sedimentary rock have a cyclic nature. This cyclic nature was caused by cyclic changes in sediment supply and the sedimentary environment. Most of these cyclic changes are caused by astronomic cycles. Short astronomic cycles can be the difference between the tides or the spring tide every two weeks. On a larger time-scale, cyclic changes in climate and sea level are caused by Milankovitch cycles: cyclic changes in the orientation and/or position of the Earth's rotational axis and orbit around the Sun. There are a number of Milankovitch cycles known, lasting between 10,000 and 200,000 years.

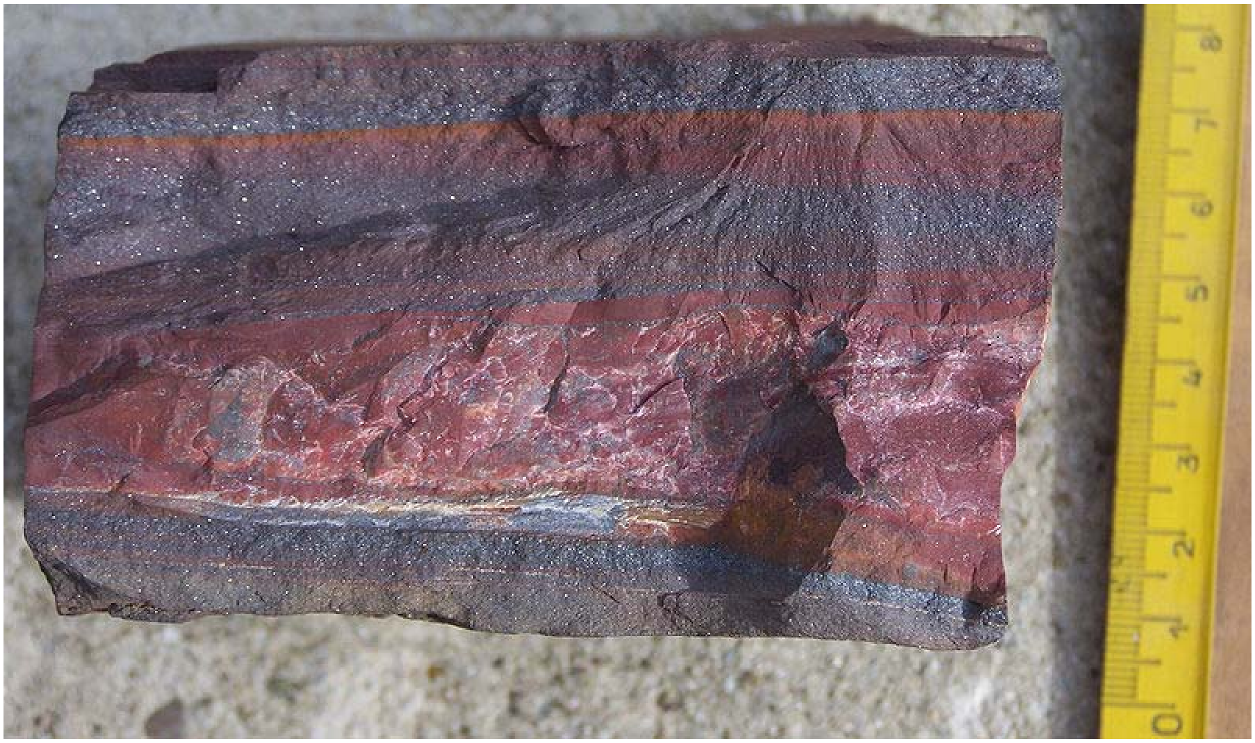
Relatively small changes in the orientation of the Earth's axis or length of the seasons can be a major influence on the Earth's climate. An example are the ice ages of the past 2.6 million years (the Quaternary period), which are assumed to have been caused by astronomic cycles. Climate change can influence the global sea level (and thus the amount of accommodation space in sedimentary basins) and sediment supply from a certain region. Eventually, small changes in astronomic parameters can cause large changes in sedimentary environment and sedimentation.

Sedimentation rates

The rate at which sediment is deposited differs depending on the location. A channel in a tidal flat can see the deposition of a few metres of sediment in one day, while on the deep ocean floor each year only a few millimetres of sediment accumulate. A distinction can be made between normal sedimentation and sedimentation caused by catastrophic processes. The latter category includes all kinds of sudden exceptional processes like mass movements, rock slides or flooding. Catastrophic processes can see the sudden deposition of a large amount of sediment at once. In some sedimentary environments, most of the total column of sedimentary rock was formed by catastrophic processes, even though the environment is usually a quiet place. Other sedimentary environments are dominated by normal, ongoing sedimentation.

In some sedimentary environments, sedimentation only occurs in some places. In a desert, for example, the wind deposits siliciclastic material (sand or silt) in some spots, or catastrophic flooding of a wadi may cause sudden deposis of large quantities of detrital material, but in most places eolian erosion dominates. The amount of sedimentary rock that forms is not only dependent on the amount of supplied material, but also on how well the material consolidates. Erosion removes most deposited sediment shortly after deposition.

Properties



A piece of a banded iron formation, a type of rock that consists of alternating layers with iron(III) oxide (red) and iron(II) oxide (grey). BIFs were mostly formed during the Precambrian, when the atmosphere wasn't yet rich in oxygen. Moories Group, Barberton Greenstone Belt, South Africa.

Color

The color of a sedimentary rock is often mostly determined by iron, an element with two major oxides: iron(II) oxide and iron(III) oxide. Iron(II) oxide only forms under anoxic circumstances and gives the rock a grey or greenish colour. Iron(III) oxide is often in the form of the mineral hematite and gives the rock a reddish to brownish colour. In arid continental climates rocks are in direct contact with the atmosphere, and oxidation is an important process, giving the rock a red or orange colour. Thick sequences of red sedimentary rocks formed in arid climates are called red beds. However, a red colour does not necessarily mean the rock formed in a continental environment or arid climate.

The presence of organic material can colour a rock black or grey. Organic material is in nature formed from dead organisms, mostly plants. Normally, such material eventually decays by oxidation or bacterial activity. Under anoxic circumstances, however, organic material cannot decay and becomes a dark sediment, rich in organic material. This, can for example, occur at the bottom of deep seas and lakes. There is little water current in such environments, so oxygen from surface water is not brought down, and the deposited sediment is normally a fine dark clay. Dark rocks rich in organic material are therefore often shales.

Texture

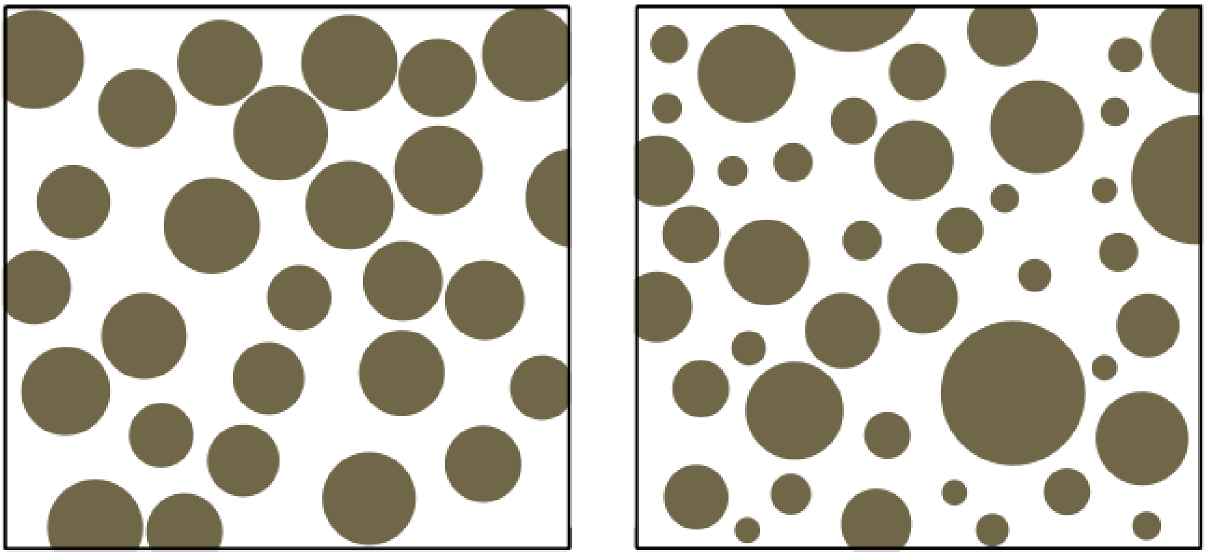


Diagram showing the difference between well-sorted (left) and poorly sorted (right) clastic rocks.

The size, form and orientation of clasts or minerals in a rock is called its texture. The texture is a small-scale property of a rock, but determined many of its large-scale properties, such as the density, porosity or permeabililty.

Clastic rocks have a 'clastic texture', which means they consist of clasts. The 3D orientation of these clasts is called the fabric of the rock. Between the clasts the rock can be composed of a matrix or a cement (the latter can consist of crystals of one or more precipitated minerals). The size and form of clasts can be used to determine the velocity and direction of current in the sedimentary environment where the rock was formed; fine, calcareous mud only settles in quiet water, while gravel and larger clasts are only deposited by rapidly moving water. The grain size of a rock is usually expressed with the Wentworth scale, though alternative scales are used sometimes. The grain size can be expressed as a diameter or a volume, and is always an average value - a rock is composed of clasts with different sizes. The statistical distribution of grain sizes is different for different rock types and is described in a property called the sorting of the rock. When all clasts are more or less of the same size, the rock is called 'well-sorted', when there is a large spread in grain size, the rock is called 'poorly sorted'.

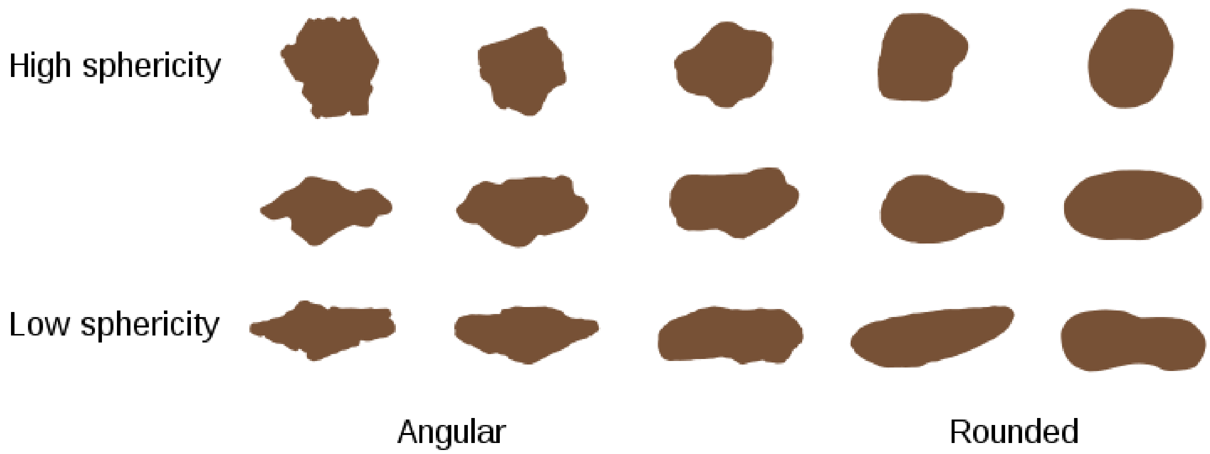


Diagram showing the influence of rounding and sphericity

The form of clasts can reflect the origin of the rock.

Coquina, a rock composed of clasts of broken shells, can only form in energetic water. The form of a clast can be described by using four parameters:

- *Surface texture* describes the amount of small-scale relief of the surface of a grain that is too small to influence the general shape.
- *rounding* describes the general smoothness of the shape of a grain.
- 'Sphericity' describes the degree to which the grain approaches a sphere.
- 'Grain form' describes the three dimensional shape of the grain.

Chemical sedimentary rocks have a non-clastic texture, consisting entirely of crystals. To describe such a texture only the average size of the crystals and the fabric are necessary.

Mineralogy

Most sedimentary rocks contain either quartz (especially siliciclastic rocks) or calcite (especially carbonate rocks). In contrast with igneous and metamorphic rocks, a sedimentary rocks usually contains very few different major minerals. However, the origin of the minerals in a sedimentary rock is often more complex than those in an igneous rock. Minerals in a sedimentary rock can have formed by precipitation during sedimentation or diagenesis. In the second case, the mineral precipitate can have grown over an older generation of cement. A complex diagenetic history can be studied by optical mineralogy, using a petrographic microscope.

Carbonate rocks dominantly consist of carbonate minerals like calcite, aragonite or dolomite. Both cement and clasts (including fossils and ooids) of a carbonate rock can consist of carbonate minerals. The mineralogy of a clastic rock is determined by the supplied material from the source area, the manner of transport to the place of deposition and the stability of a particular mineral. The stability of the major rock forming minerals (their resistance to weathering) is expressed by

Bowen's reaction series. In this series, quartz is most stable, followed by feldspar, micas, and other less stable minerals that are only present when little weathering has occurred. The amount of weathering depends mainly on the distance to the source area, the local climate and the time it took for the sediment to be transported there. In most sedimentary rocks, mica, feldspar and less stable minerals have reacted to clay minerals like kaolinite, illite or smectite.

Primary sedimentary structures

Cross-bedding in a fluviatile sandstone, Middle Old Red Sandstone (Devonian) on Bressay, Shetland Islands.

A flute cast, a type of sole marking, from the Book Cliffs of Utah

Ripple marks formed by a current in a sandstone that was later tilted. Location: Haßberge, Bavaria.

Structures in sedimentary rocks can be divided in 'primary' structures (formed during deposition) and 'secondary' structures (formed after deposition). Unlike textures, structures are always large-scale features that can easily be studied in the field. Sedimentary structures can tell something about the sedimentary environment or can serve to tell which side originally faced up where tectonics have tilted or overturned sedimentary layers.

Sedimentary rocks are laid down in layers called beds or strata. A bed is defined as a layer of rock that has a uniform lithology and texture. Beds form by the deposition of layers of sediment on top of each other. The sequence of beds that characterizes sedimentary rocks is called bedding. Single beds can be a couple of centimetres to several meters thick. Finer, less pronounced layers are called laminae and the structure it forms in a rock is called lamination. Laminae are usually less than a few centimetres thick. Though bedding and lamination are often originally horizontal in nature, this is not always the case. In some environments, beds are deposited at a (usually small) angle. Sometimes multiple sets of layers with different orientations exist in the same rock, a structure called cross-bedding. Cross-bedding forms when small-scale erosion occurs during deposition, cutting off part of the beds. Newer beds then form at an angle to older ones.

The opposite of cross-bedding is parallel lamination, where all sedimentary layering is parallel. With laminations, differences are generally caused by cyclic changes in the sediment supply, caused for example by seasonal changes in rainfall, temperature or biochemical activity. Laminae that represent seasonal changes (like tree rings) are called varves. Some rocks have no lamination at all, their structural character is called massive bedding.

Graded bedding is a structure where beds with a smaller grain size occur on top of beds with larger grains. This structure forms when fast flowing water stops flowing. Larger, heavier clasts in suspension settle first, then smaller clasts. Though graded bedding can form in many different environments, it is characteristic for turbidity currents.

The bedform (the surface of a particular bed) can be indicative for a particular sedimentary environment too. Examples of bed forms are sole markings and ripple marks. Sole markings, such as tool marks and flute casts, are grove dug into a sedimentary layer that are preserved. These are often elongated structures and can be used to establish the direction of the flow during deposition.

Ripple marks also form in flowing water. There are two types: asymmetric wave ripples and symmetric current ripples. Environments where the current is in one direction, such as rivers, produce asymmetric ripples. The longer flank of such ripples is oriented opposite to the direction of the current. Wave ripples occur in environments where currents occur in all directions, such as tidal flats.

Another type of bed form are mud cracks, caused by the dehydration of sediment that occasionally comes above the water surface. Such structures are commonly found at tidal flats or point bars along rivers.

Fossils

Fossil-rich layers in a sedimentary rock, Año Nuevo State Reserve, California

Sedimentary rocks are the only type of rock that can contain fossils, the remains or imprints of dead organisms. In nature, dead organisms are usually quickly removed by scavengers, bacteria, rotting and erosion. In some exceptional circumstances a carcass is fossilized because these natural processes are unable to work. The chance of fossilisation is higher when the sedimentation rate is high (so that a carcass is quickly buried), in anoxic environments (where little bacterial activity exists) or when the organism had a particularly hard skeleton. Larger, well-preserved fossils are relatively rare. Most sedimentary rocks contains fossils, though with many the fact only becomes apparent when studied under a microscope (microfossils) or with a loupe.

Burrows in a turbidite, made by crustaceans. San Vincente Formation (early Eocene) of the Ainsa Basin, southern foreland of the Pyrenees.

Fossils can both be the direct remains or imprints of organisms and their skeletons. Most commonly preserved are the harder parts of organisms such as bones, shells, woody tissue of plants. Soft tissue has a much smaller chance of being preserved and fossilized and soft tissue of animals older than 40 million years is very rare. Imprints of organisms made while still alive are called trace fossils. Examples are burrows, foot prints, etc.

Being part of a sedimentary rock, fossils undergo the same diagenetic processes as the rock. A shell consisting of calcite can for example dissolve, while a cement of silica then fills the cavity. In the same way, precipitating minerals can fill cavities formerly occupied by blood vessels, vascular tissue or other soft tissues. This preserves the form of the organism but changes the chemical composition, a process called permineralisation. The most common minerals in permineralisation cements are carbonates (especially calcite), forms of amorphous silica (chalcedony, flint, chert) and pyrite. In the case of silica cements, the process is called lithification.

At high pressure and temperature, the organic material of a dead organism undergoes chemical reactions in which volatiles like water and carbon dioxide are expulsed. The fossil, in the end, consists of a thin layer of pure carbon or its mineralized form, graphite. This form of fossilisation is called carbonisation. It is particularly important for plant fossils. The same process is responsible for the formation of fossil fuels like lignite or coal.

Stratigraphy

The Permian through Jurassic stratigraphy of the Colorado Plateau area of southeastern Utah that makes up much of the famous prominent rock formations in protected areas such as Capitol Reef National Park and Canyonlands National Park. From top to bottom: Rounded tan domes of the Navajo Sandstone, layered red Kayenta Formation, cliff-forming, vertically jointed, red Wingate Sandstone, slope-forming, purplish Chinle Formation, layered, lighter-red Moenkopi Formation, and white, layered Cutler Formation sandstone. Picture from Glen Canyon National Recreation Area, Utah.

That new rock layers are above older rock layers is stated in the principle of superposition. There are usually some gaps in the sequence called unconformities. These represent periods where no new sediments were laid down, or when earlier sedimentary layers raised above sea level and eroded away.

Sedimentary rocks contain important information about the history of the Earth. They contain fossils, the preserved remains of ancient plants and animals. Coal is considered a type of sedimentary rock. The composition of sediments provides us with clues as to the original rock. Differences between successive layers indicate changes to the environment over time. Sedimentary rocks can contain fossils because, unlike most igneous and metamorphic rocks, they form at temperatures and pressures that do not destroy fossil remains.

2. Metamorphic rock

Quartzite, a form of metamorphic rock, from the Museum of Geology at University of Tartu collection

Metamorphic rock is the result of the transformation of an existing rock type, the *protolith*, in a process called metamorphism, which means "change in form". The protolith is subjected to heat and pressure (temperatures greater than 150 to 200 °C and pressures of 1500 bars) causing profound physical and/or chemical change. The protolith may be sedimentary rock, igneous rock or another older metamorphic rock. Metamorphic rocks make up a large part of the Earth's crust and are classified by texture and by chemical and mineral assemblage (metamorphic facies). They may be formed simply by being deep beneath the Earth's surface, subjected to high temperatures and the great pressure of the rock layers above it. They can form from tectonic processes such as continental collisions, which cause horizontal pressure, friction and distortion. They are also formed when rock is heated up by the intrusion of hot molten rock called magma from the Earth's interior. The study of metamorphic rocks (now exposed at the Earth's surface following erosion and uplift) provides us with information about the temperatures and pressures that occur at great depths within the Earth's crust.

Some examples of metamorphic rocks are gneiss, slate, marble, schist, and quartzite.

Metamorphic minerals

Metamorphic minerals are those that form only at the high temperatures and pressures associated with the process of metamorphism. These minerals, known as index minerals, include sillimanite, kyanite, staurolite, andalusite, and some garnet.

Other minerals, such as olivines, pyroxenes, amphiboles, micas, feldspars, and quartz, may be found in metamorphic rocks, but are not necessarily the result of the process of metamorphism. These minerals formed during the crystallization of igneous rocks. They are stable at high temperatures and pressures and may remain chemically unchanged during the metamorphic process. However, all minerals are stable only within certain limits, and the presence of some minerals in metamorphic rocks indicates the approximate temperatures and pressures at which they formed.

The change in the particle size of the rock during the process of metamorphism is called recrystallization. For instance, the small calcite crystals in the sedimentary rock limestone change into larger crystals in the metamorphic rock marble, or in metamorphosed sandstone, recrystallization of the original quartz sand grains results in very compact quartzite, in which the often larger quartz crystals are interlocked. Both high temperatures and pressures contribute to recrystallization. High temperatures allow the atoms and ions in solid crystals to migrate, thus reorganizing the crystals, while high pressures cause solution of the crystals within the rock at their point of contact.

Foliation

Folded foliation in a metamorphic rock from near Geirangerfjord, Norway

The layering within metamorphic rocks is called *foliation* (derived from the Latin word *folia*, meaning "leaves"), and it occurs when a rock is being shortened along one axis during recrystallization. This causes the platy or elongated crystals of minerals, such as mica and chlorite, to become rotated such that their long axes are perpendicular to the orientation of shortening. This results in a banded, or foliated, rock, with the bands showing the colors of the minerals that formed them.

Textures are separated into foliated and non-foliated categories. Foliated rock is a product of differential stress that deforms the rock in one plane, sometimes creating a plane of cleavage. For example, slate is a foliated metamorphic rock, originating from shale. Non-foliated rock does not have planar patterns of strain.

Rocks that were subjected to uniform pressure from all sides, or those that lack minerals with distinctive growth habits, will not be foliated. Slate is an example of a very fine-grained, foliated metamorphic rock, while phyllite is medium, schist coarse, and gneiss very coarse-grained. Marble is generally not foliated, which allows its use as a material for sculpture and architecture.

Another important mechanism of metamorphism is that of chemical reactions that occur between minerals without them melting. In the process atoms are exchanged between the minerals, and thus new minerals are formed. Many complex high-temperature reactions may take place, and each mineral assemblage produced provides us with a clue as to the temperatures and pressures at the time of metamorphism.

Metasomatism is the drastic change in the bulk chemical composition of a rock that often occurs during the processes of metamorphism. It is due to the introduction of chemicals from other surrounding rocks. Water may transport these chemicals rapidly over great distances. Because of the role played by water, metamorphic rocks generally contain many elements absent from the original rock, and lack some that originally were present. Still, the introduction of new chemicals is not necessary for recrystallization to occur.

Types of metamorphism

Contact metamorphism

A contact metamorphic rock made of interlayered calcite and serpentine from the Precambrian of Canada. Once thought to be a fossil called *Eozoön canadense*. Scale in mm.

Contact metamorphism is the name given to the changes that take place when magma is injected into the surrounding solid rock (country rock). The changes that occur are greatest wherever the magma comes into contact with the rock because the temperatures are highest at this boundary and decrease with distance from it. Around the igneous rock that forms from the cooling magma is a metamorphosed zone called a *contact metamorphism aureole*. Aureoles may show all degrees of metamorphism from the contact area to unmetamorphosed (unchanged) country rock some distance away. The formation of important ore minerals may occur by the process of metasomatism at or near the contact zone.

When a rock is contact altered by an igneous intrusion it very frequently becomes more indurated, and more coarsely crystalline. Many altered rocks of this type were formerly called hornstones, and the term *hornfels* is often used by geologists to signify those fine grained, compact, non-foliated products of contact metamorphism. A shale may become a dark argillaceous hornfels, full of tiny plates of brownish biotite; a marl or impure limestone may change to a grey, yellow or greenish lime-silicate-hornfels or siliceous marble, tough and splintery, with abundant augite, garnet, wollastonite and other minerals in which calcite is an important component. A diabase or andesite may become a diabase hornfels or andesite hornfels with development of new hornblende and biotite and a partial recrystallization of the original feldspar. Chert or flint may become a finely crystalline quartz rock; sandstones lose their clastic structure and are converted into a mosaic of small close-fitting grains of quartz in a metamorphic rock called quartzite.

If the rock was originally banded or foliated (as, for example, a laminated sandstone or a foliated calc-schist) this character may not be obliterated, and a banded hornfels is the product; fossils even may have their shapes preserved, though entirely recrystallized, and in many contact-altered lavas the vesicles are still visible, though their contents have usually entered into new combinations to form minerals that were not originally present. The minute structures, however, disappear, often completely, if the thermal alteration is very profound; thus small grains of quartz in a shale are lost or blend with the surrounding particles of clay, and the fine ground-mass of lavas is entirely reconstructed.

By recrystallization in this manner peculiar rocks of very distinct types are often produced. Thus shales may pass into cordierite rocks, or may show large crystals of andalusite (and chiastolite), staurolite, garnet, kyanite and sillimanite, all derived from the aluminous content of the original shale. A considerable amount of mica (both muscovite and biotite) is often simultaneously formed, and the resulting product has a close resemblance to many kinds of schist. Limestones, if pure, are often turned into coarsely crystalline marbles; but if there was an admixture of clay or sand in the original rock such minerals as garnet, epidote, idocrase, wollastonite, will be present. Sandstones when greatly heated may change into coarse quartzites composed of large clear grains of quartz. These more intense stages of alteration are not so commonly seen in igneous rocks, because their minerals, being formed at high temperatures, are not so easily transformed or recrystallized.

In a few cases rocks are fused and in the dark glassy product minute crystals of spinel, sillimanite and cordierite may separate out. Shales are occasionally thus altered by basalt dikes, and feldspathic sandstones may be completely vitrified. Similar changes may be induced in shales by the burning of coal seams or even by an ordinary furnace.

There is also a tendency for metasomatism between the igneous magma and sedimentary country rock, whereby the chemicals in each are exchanged or introduced into the other. Granites may absorb fragments of shale or pieces of basalt. In that case, hybrid rocks called skarn arise, which don't have the characteristics of normal igneous or sedimentary rocks. Sometimes an invading granite magma permeates the rocks around, filling their joints and planes of bedding, etc., with threads of quartz and feldspar. This is very exceptional but instances of it are known and it may take place on a large scale.

Regional metamorphism

Mississippian marble in Big Cottonwood Canyon, Wasatch Mountains, Utah

Regional metamorphism is the name given to changes in great masses of rock over a wide area. Rocks can be metamorphosed simply by being at great depths below the Earth's surface, subjected to high temperatures and the great pressure caused by the immense weight of the rock layers above. Much of the lower continental crust is metamorphic, except for recent igneous intrusions. Horizontal tectonic movements such as the collision of continents create orogenic belts, and cause high temperatures, pressures and deformation in the rocks along these belts. If the metamorphosed rocks are later uplifted and exposed by erosion, they may occur in long belts or other large areas at the surface. The process of metamorphism may have destroyed the original features that could have revealed the rock's previous history. Recrystallization of the rock will destroy the textures and fossils present in sedimentary rocks. Metasomatism will change the original composition.

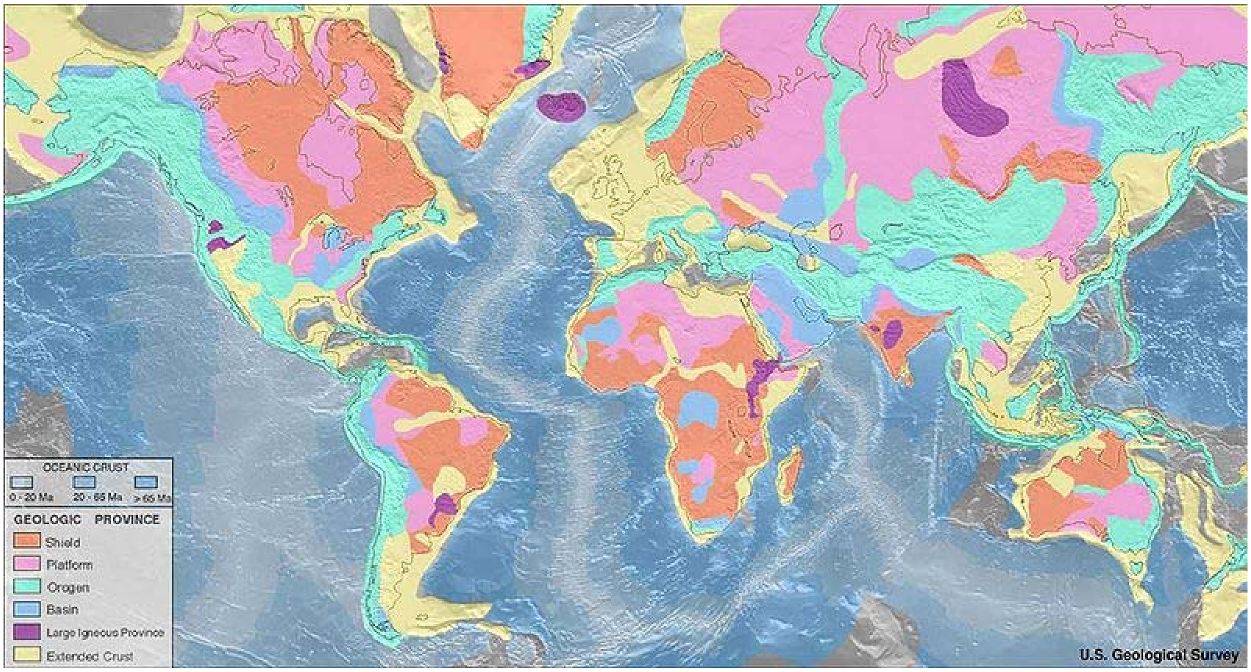
Regional metamorphism tends to make the rock more indurated and at the same time to give it a foliated, shistose or gneissic texture, consisting of a planar arrangement of the minerals, so that platy or prismatic minerals like mica and hornblende have their longest axes arranged parallel to one another. For that reason many of these rocks split readily in one direction along mica-bearing zones (schists). In gneisses, minerals also tend to be segregated into bands; thus there are seams of quartz and of mica in a mica schist, very thin, but consisting essentially of one mineral. Along the mineral layers composed of soft or fissile minerals the rocks will split most readily, and the freshly split specimens will appear to be faced or coated with this mineral; for example, a piece of mica schist looked at facewise might be supposed to consist entirely of shining scales of mica. On the edge of the specimens, however, the white folia of granular quartz will be visible. In gneisses these alternating folia are sometimes thicker and less regular than in schists, but most importantly less micaceous; they may be lenticular, dying out rapidly. Gneisses also, as a rule, contain more feldspar than schists do, and they are tougher and less fissile. Contortion or crumbling of the foliation is by no means uncommon, and then the splitting faces are undulose or puckered. Schistosity and gneissic banding (the two main types of foliation) are formed by directed pressure at elevated temperature, and to interstitial movement, or internal flow arranging the mineral particles while they are crystallizing in that directed pressure field.

Rocks that were originally sedimentary and rocks that were undoubtedly igneous convert into schists and gneisses. If originally of similar composition they may be very difficult to distinguish from one another if the metamorphism has been great. A quartz-porphyry, for example, and a fine feldspathic sandstone, may both the converted into a grey or pink mica-schist.

Metamorphic rock textures

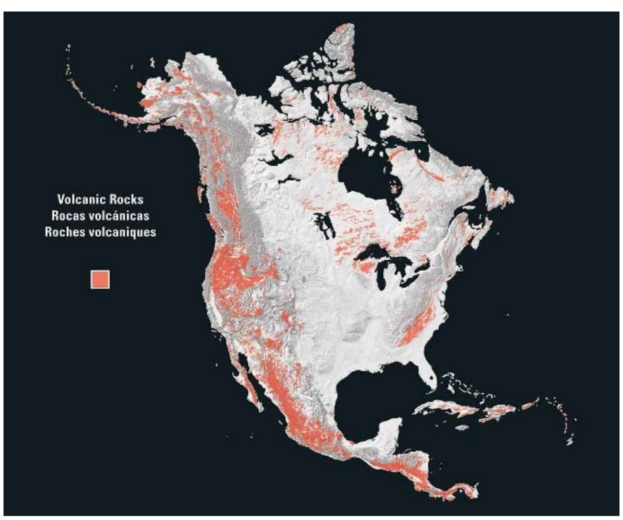
The five basic metamorphic textures with typical rock types are slaty (includes slate and phyllite) (the foliation is called "slaty cleavage"), "schistose" (includes schist) (the foliation is called ("schistosity"), gneissose (gneiss) (the foliation is called "gneissosity"), granoblastic (includes granulite, some marbles and quartzite), and hornfelsic (includes hornfels and skarn).

3. Igneous rock

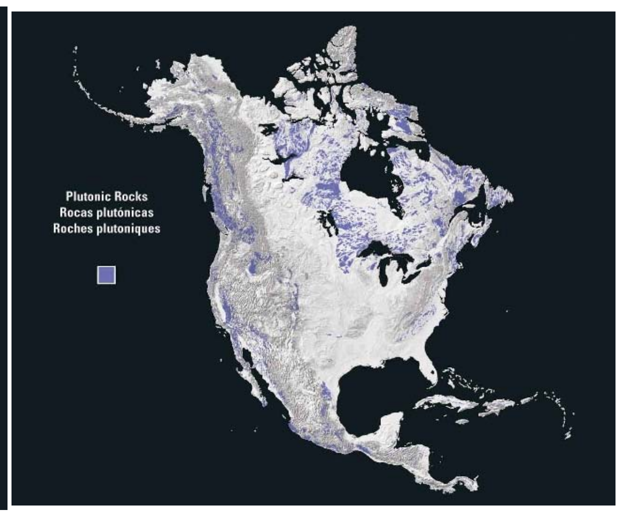


Geologic provinces of the world (USGS)
Shield Platform Orogen Basin Large igneous province Extended crust

Oceanic crust: 0–20 Ma 20–65 Ma >65 Ma



Volcanic rock in North America.



Plutonic rock in North America.

Igneous rock (derived from the Latin word *igneus* meaning of fire, from *ignis* meaning fire) is one of the three main rock types, the others being sedimentary and metamorphic rock. Igneous rock is formed through the cooling and solidification of magma or lava. Igneous rock may form with or without crystallization, either below the surface as intrusive (plutonic) rocks or on the surface as extrusive (volcanic) rocks. This magma can be derived from partial melts of pre-existing rocks in either a planet's mantle or crust. Typically, the melting is caused by one or more of three processes: an increase in temperature, a decrease in pressure, or a change in composition. Over 700 types of igneous rocks have been described, most of them having formed beneath the surface of Earth's crust. These have diverse properties, depending on their composition and how they were formed.

Geological significance

The upper 16 kilometres (10 mi) of Earth's crust is composed of approximately 95% igneous rocks with only a thin, widespread covering of sedimentary and metamorphic rocks.

Igneous rocks are geologically important because:

- their minerals and global chemistry give information about the composition of the mantle, from which some igneous rocks are extracted, and the temperature and pressure conditions that allowed this extraction, and/or of other pre-existing rock that melted;
- their absolute ages can be obtained from various forms of radiometric dating and thus can be compared to adjacent geological strata, allowing a time sequence of events;
- their features are usually characteristic of a specific tectonic environment, allowing tectonic reconstitutions ;
- in some special circumstances they host important mineral deposits (ores): for example, tungsten, tin, and uranium are commonly associated with granites and diorites, whereas ores of chromium and platinum are commonly associated with gabbros.

Morphology and setting

In terms of modes of occurrence, igneous rocks can be either intrusive (plutonic), extrusive (volcanic) or hypabyssal.

Intrusive igneous rocks

Close-up of granite (an intrusive igneous rock) exposed in Chennai, India.

Intrusive igneous rocks are formed from magma that cools and solidifies within the crust of a planet. Surrounded by pre-existing rock (called country rock), the magma cools slowly, and as a result these rocks are coarse grained. The mineral grains in such rocks can generally be identified with the naked eye. Intrusive rocks can also be classified according to the shape and size of the intrusive body and its relation to the other formations into which it intrudes. Typical intrusive formations are batholiths, stocks, laccoliths, sills and dikes.

The central cores of major mountain ranges consist of intrusive igneous rocks, usually granite. When exposed by erosion, these cores (called *batholiths*) may occupy huge areas of the Earth's surface.

Coarse grained intrusive igneous rocks which form at depth within the crust are termed as *abyssal*; intrusive igneous rocks which form near the surface are termed *hypabyssal*.

Extrusive igneous rocks

Basalt (an extrusive igneous rock in this case); light coloured tracks show the direction of lava flow

Extrusive igneous rocks are formed at the crust's surface as a result of the partial melting of rocks within the mantle and crust. Extrusive Igneous rocks cool and solidify quicker than intrusive igneous rocks. Since the rocks cool very quickly they are fine grained.

The melted rock, with or without suspended crystals and gas bubbles, is called magma. Magma rises because it is less dense than the rock from which it was created. When it reaches the surface, magma extruded onto the surface either beneath water or air, is called lava. Eruptions of volcanoes into air are termed *subaerial* whereas those occurring underneath the ocean are termed *submarine*. Black smokers and mid-ocean ridge basalt are examples of submarine volcanic activity.

The volume of extrusive rock erupted annually by volcanoes varies with plate tectonic setting. Extrusive rock is produced in the following proportions:

- divergent boundary: 73%

- convergent boundary (subduction zone): 15%
- hotspot: 12%.

Magma which erupts from a volcano behaves according to its viscosity, determined by temperature, composition, and crystal content. High-temperature magma, most of which is basaltic in composition, behaves in a manner similar to thick oil and, as it cools, treacle. Long, thin basalt flows with pahoehoe surfaces are common. Intermediate composition magma such as andesite tends to form cinder cones of intermingled ash, tuff and lava, and may have viscosity similar to thick, cold molasses or even rubber when erupted. Felsic magma such as rhyolite is usually erupted at low temperature and is up to 10,000 times as viscous as basalt. Volcanoes with rhyolitic magma commonly erupt explosively, and rhyolitic lava flows typically are of limited extent and have steep margins, because the magma is so viscous.

Felsic and intermediate magmas that erupt often do so violently, with explosions driven by release of dissolved gases — typically water but also carbon dioxide. Explosively erupted pyroclastic material is called tephra and includes tuff, agglomerate and ignimbrite. Fine volcanic ash is also erupted and forms ash tuff deposits which can often cover vast areas.

Because lava cools and crystallizes rapidly, it is fine grained. If the cooling has been so rapid as to prevent the formation of even small crystals after extrusion, the resulting rock may be mostly glass (such as the rock obsidian). If the cooling of the lava happened slowly, the rocks would be coarse-grained.

Because the minerals are mostly fine-grained, it is much more difficult to distinguish between the different types of extrusive igneous rocks than between different types of intrusive igneous rocks. Generally, the mineral constituents of fine-grained extrusive igneous rocks can only be determined by examination of thin sections of the rock under a microscope, so only an approximate classification can usually be made in the field.

Hypabyssal igneous rocks

Hypabyssal igneous rocks are formed at a depth in between the plutonic and volcanic rocks. Hypabyssal rocks are less common than plutonic or volcanic rocks and do often form dikes, sills or laccoliths.

Classification of Igneous Rock

Pegmatite

Pegmatite with blue corundum crystals

Pegmatite containing lepidolite, tourmaline, and quartz from the White Elephant Mine in the Black Hills, South Dakota.

A **pegmatite** is a very coarse-grained, intrusive igneous rock composed of interlocking grains usually larger than 2.5 cm in size; such rocks are referred to as *pegmatitic*.

Most pegmatites are composed of quartz, feldspar and mica; in essence a granite. Rarer intermediate composition and mafic pegmatites containing amphibole, Ca-plagioclase feldspar, pyroxene and other minerals are known, found in recrystallised zones and apophyses associated with large layered intrusions.

Crystal size is the most striking feature of pegmatites, with crystals usually over 5 cm in size. Individual crystals over 10 meters across have been found, and the world's largest crystal was found within a pegmatite.

Similarly, crystal texture and form within pegmatitic rock may be taken to extreme size and perfection. Feldspar within a pegmatite may display exaggerated and perfect twinning, exsolution lamellae, and when affected by hydrous crystallization, macroscale graphic texture is known, with feldspar and quartz intergrown. Perthite feldspar within a pegmatite often shows gigantic perthitic texture visible to the naked eye.

Petrology

Crystal growth rates in pegmatite must be incredibly fast to allow gigantic crystals to grow within the confines and pressures of the Earth's crust. For this reason, the consensus on pegmatitic growth mechanisms involves a combination of the following processes;

- Low rates of nucleation of crystals coupled with high diffusivity to force growth of a few large crystals instead of many smaller crystals
- High vapor and water pressure, to assist in the enhancement of conditions of diffusivity
- High concentrations of fluxing elements such as boron and lithium which lower the temperature of solidification within the magma or vapor
- Low thermal gradients coupled with a high wall rock temperature, explaining the preponderance for pegmatite to occur only within greenschist metamorphic terranes

Despite this consensus on likely chemical, thermal and compositional conditions required to promote pegmatite growth there are three main theories behind pegmatite formation;

1. Metamorphic; pegmatite fluids are created by devolatilisation (dewatering) of metamorphic rocks, particularly felsic gneiss, to liberate the right constituents and water, at the right temperature
2. Magmatic; pegmatites tend to occur in the aureoles of granites in most cases, and are usually granitic in character, often closely matching the compositions of nearby granites. Pegmatites thus represent exsolved granitic material which crystallises in the country rocks
3. Metasomatic; pegmatite, in a few cases, could be explained by the action of hot alteration fluids upon a rock mass, with bulk chemical and textural change.

Metasomatism is currently not well favored as a mechanism for pegmatite formation and it is likely that metamorphism and magmatism are both contributors toward the conditions necessary for pegmatite genesis.

Mineralogy

Pegmatitic granite, Rock Creek Canyon, eastern Sierra Nevada, California. Note pink potassium feldspars and cumulate-filled chamber.

The mineralogy of a pegmatite is in all cases dominated by some form of feldspar, often with mica and usually with quartz, being altogether "granitic" in character. Beyond that, pegmatite may include most minerals associated with granite and granite-associated hydrothermal systems, granite-associated mineralisation styles, for example greisens, and somewhat with skarn associated mineralisation.

It is however impossible to quantify the mineralogy of pegmatite in simple terms because of their varied mineralogy and difficulty in estimating the modal abundance of mineral species which are of only a trace amount. This is because of the difficulty in counting and sampling mineral grains in a rock which may have crystals from centimeters to meters across.

Garnet, commonly almandine or grossular, is a common mineral within pegmatites intruding mafic and carbonate-bearing sequences. Pegmatites associated with granitic domes within the Archaean Yilgarn Craton intruding ultramafic and mafic rocks contain red, orange and brown almandine garnet.

Tantalum and niobium minerals (columbite, tantalite, niobite) are found in association with spodumene, lepidolite, tourmaline, cassiterite in the massive Greenbushes Pegmatite in the Yilgarn Craton of Western Australia, considered a typical metamorphic pegmatite unassociated with granite.

Geochemistry

Pegmatite is difficult to sample representatively due to the large size of the constituent mineral crystals. Often, bulk samples of some 50–60 kg of rock must be crushed to obtain a meaningful and repeatable result. Hence, pegmatite is often characterised by sampling the individual minerals which comprise the pegmatite, and comparisons are made according to mineral chemistry.

Geochemically, pegmatites typically have major element compositions approximating "granite", however, when found in association with granitic plutons it is likely that a pegmatite dike will have a different trace element composition with greater enrichment in large-ion lithophile (incompatible) elements, boron, beryllium, aluminium, potassium and lithium, uranium, thorium, cesium, et cetera.

Occasionally, enrichment in the unusual trace elements will result in crystallisation of equally unusual and rare minerals such as beryl, tourmaline, columbite, tantalite, zinnwaldite and so forth. In most cases, there is no particular *genetic* significance to the presence of rare mineralogy within a pegmatite, however it is possible to see some causative and genetic links between, say, tourmaline-bearing granite dikes and tourmaline-bearing pegmatites within the area of influence of a composite granite intrusion (Mount Isa Inlier, Queensland, Australia).

Economic importance

Pegmatites are important because they often contain rare earth minerals and gemstones, such as aquamarine, tourmaline, topaz, fluorite, apatite and corundum, often along with tin and tungsten minerals, among others. For example, beautiful crystals of aquamarines and topaz can be found in pegmatites in the mountains of Colorado and Idaho.

Pegmatites are the primary source of lithium either as spodumene, lithiophyllite or usually from lepidolite (Li-mica). The only used source for caesium is also a mineral from a zoned pegmatite, pollucite. The majority of the world's beryllium is sourced from non-gem quality beryl within pegmatite. Tantalum, niobium, rare-earth elements are sourced from a few pegmatites worldwide, notably the Greenbushes Pegmatite. Bismuth, molybdenum and tin have been won from pegmatite, but this is not yet an important source of these metals.

Nomenclature

Pegmatites can be classified according to the elements or mineral of interest, for instance "lithian pegmatite" to describe a Li-bearing or Li-mineral bearing pegmatite, or "boron pegmatite" for those containing tourmaline.

There is often no meaningful way to distinguish pegmatites according to chemistry due to the difficulty of obtaining a representative sample, but often groups of pegmatites can be distinguished on contact textures, orientation, accessory minerals and timing. These may be named formally or informally as a class of intrusive rock or within a larger igneous association.

While difficult to be certain of derivation of pegmatite in the strictest sense, often pegmatites are referred to as "metamorphic", "granitic" or "metasomatic", based on the interpretations of the investigating geologist.

Occurrence

Worldwide, notable pegmatite occurrences are within the major cratons, and within greenschist-facies metamorphic belts. However, pegmatite localities are only well recorded when economic mineralisation is found.

Within the metamorphic belts, pegmatite tends to concentrate around granitic bodies within zones of low mean strain and within zones of extension, for example within the strain shadow of a large rigid granite body. Similarly, pegmatite is often found within the contact zone of granite, transitional with some greisens, as a late-stage magmatic-hydrothermal effect of syn-metamorphic granitic magmatism. Some skarns associated with granites also tend to host pegmatites.

Aplite and porphyry dikes and veins may intrude pegmatites and wall rocks adjacent to intrusions, creating a confused sequence of felsic intrusive apophyses (thin branches or offshoots of igneous bodies) within the aureole of some granites.

Phaneritic

Close-up of granite exposed in Chennai, India.

Phaneritic is a term usually used to refer to igneous rock grain size. It means that the size of matrix grains in the rock are large enough to be distinguished with the unaided eye as opposed to aphanitic (which is too small to see with the naked eye). This texture forms by slow cooling of magma deep underground in the plutonic environment. It may also be applied to metamorphic rocks with the same definition.

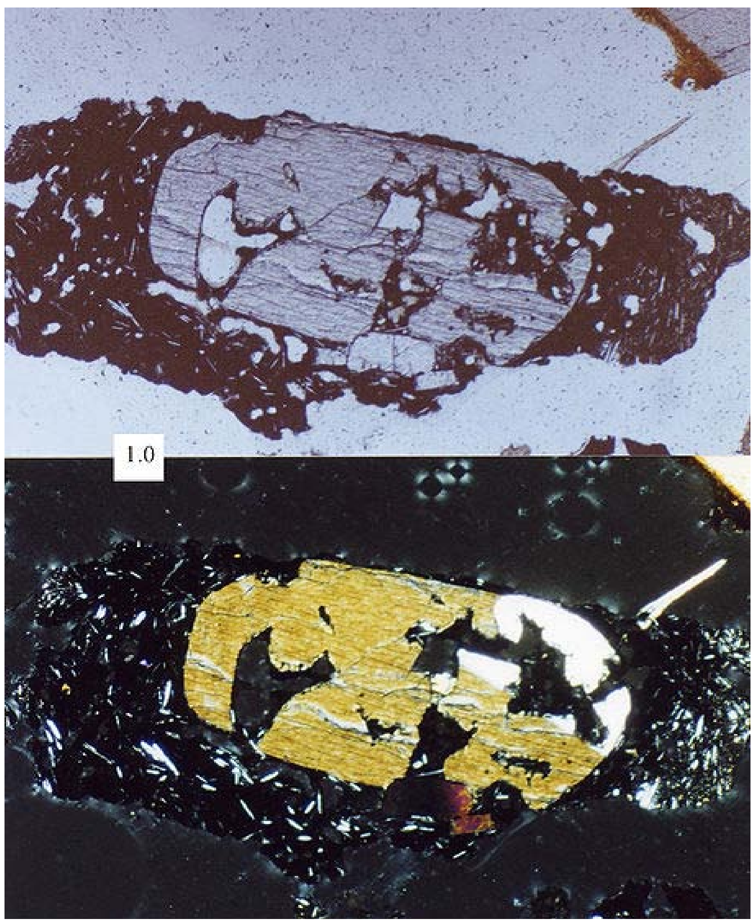
Aphanite

Aphanite, or **aphanitic** as an adjective, (from the Greek αφανης, invisible) is a name given to certain igneous rocks which are so fine grained that their component mineral crystals are not detected by the unaided eye (as opposed to phaneritic igneous rocks, where the minerals are visible to the unaided eye). This texture results from rapid cooling in volcanic or hypabyssal (shallow subsurface) environments. As a rule, the texture of these rocks are not quite the same volcanic glass (e.g. obsidian), with volcanic glass being even finer grained (or more accurately, non-crystalline) than aphanitic rocks, and having a glass-like appearance.

Aphanites are commonly porphyritic, having large crystals embedded in the fine groundmass or matrix. The large inclusions are called phenocrysts.

They consist essentially of very fine grained minerals, such as plagioclase feldspar, with hornblende or augite, and may contain also biotite, quartz, and orthoclase.

Porphyritic



A porphyritic volcanic sand grain, as seen under the petrographic microscope. The large grain in the middle is of a much different size class than the small needle-like crystals around it. Scale box in millimeters.

Andesite porphyry from summit of O'Leary Peak. This is an extrusive porphyritic rock, as the pink (and black) phenocrysts are clearly visible, in contrast the grey groundmass with its microscopic crystals.

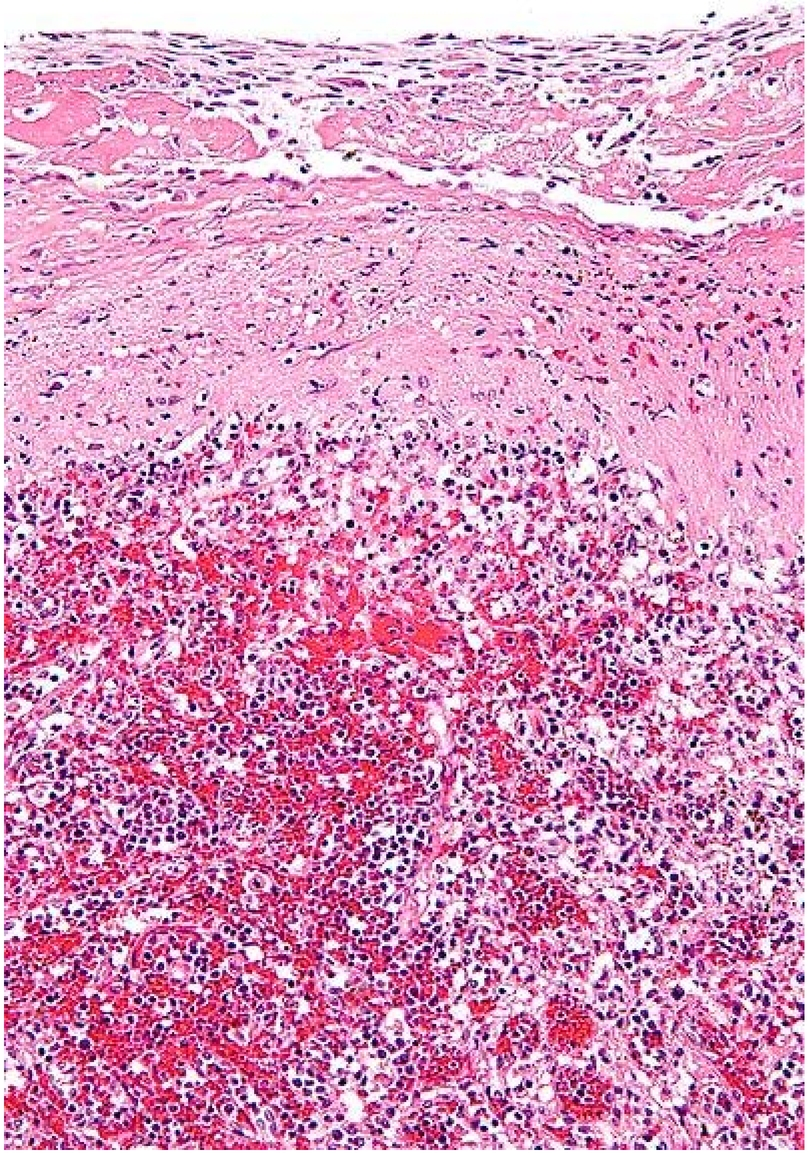
Porphyritic texture in a granite. This is an intrusive porphyritic rock. The white, square feldspar phenocrysts are much larger than crystals in the surrounding matrix; eastern Sierra Nevada, Rock Creek Canyon, California.

Porphyritic is an adjective used in geology, specifically for igneous rocks, for a rock that has a distinct difference in the size of the crystals, with at least one group of crystals obviously larger than another group. Porphyritic rocks may be aphanites or extrusive, with large crystals or phenocrysts floating in a fine-grained groundmass of non-visible crystals, as in a porphyritic basalt, or phanerites or intrusive, with individual crystals of the groundmass may be easily distinguished with the eye, but one group of crystals clearly much bigger than the rest, as in a porphyritic granite. Most types of igneous rocks may display some degree of porphyritic texture. One main type of rock that has a porphyritic texture are porphyry, though not all porphyritic rocks are porphyries.

Formation

Porphyritic rocks are formed when a column of rising magma is cooled in two stages. In the first stage, the magma is cooled slowly deep in the crust, creating the large crystal grains, with a diameter of 2 mm or more. In the final stage, the magma is cooled rapidly at relatively shallow depth or as it erupts from a volcano, creating small grains that are usually invisible to the unaided eye.

Hyaline



Micrograph of spleen with hyaline deposition (pink material - top of image) in association with inflammation (hyaloserositis). H&E stain.

The term **hyaline** (from Greek: ὕαλος 'glassy') literally refers to a substance with a glass-like appearance.

In common medical histopathological usage, hyaline is a substance with a glassy, pink appearance after haematoxylin and eosin staining—most often an acellular, proteinaceous material. Hyaline cartilage is the clear, shiny cartilage of articular joints.

In other scientific fields such as ichthyology and entomology, hyaline may refer to a specific type of colorless and transparent substance.

Pyroclastic rock

USGS scientist examines pumice blocks at the edge of a pyroclastic flow from Mount St. Helens

Rocks from the Bishop tuff, uncompressed with pumice on left; compressed with fiamme on right

Pyroclastic rocks or **pyroclastics** (derived from the Greek *πῦρ*, meaning fire; and *κλαστός*, meaning broken) are clastic rocks composed solely or primarily of volcanic materials. Where the volcanic material has been transported and reworked through mechanical action, such as by wind or water, these rocks are termed ***volcaniclastic***. Commonly associated with explosive volcanic activity–such as Plinian or krakatoan eruption styles, or phreatomagmatic eruptions–pyroclastic deposits are commonly formed from airborne ash, lapilli and bombs or blocks ejected from the volcano itself, mixed in with shattered country rock.

Pyroclastic rocks may be composed of a large range of clast sizes; from the largest agglomerates, to very fine ashes and tuffs. Pyroclasts of different sizes are classified as volcanic bombs, lapilli and volcanic ash. Ash is considered to be pyroclastic because it is a fine dust made up of volcanic rock. One of the most spectacular forms of pyroclastic deposit are the ignimbrites, deposits formed by the high-temperature gas and ash mix of a pyroclastic flow event.

Three modes of transport can be distinguished: pyroclastic flow, pyroclastic surge, and pyroclastic fall. During Plinian eruptions, pumice and ash are formed when silicic magma is fragmented in the volcanic conduit, because of decompression and the growth of bubbles.

Pyroclasts are then entrained in a buoyant eruption plume which can rise several kilometers into the air and cause aviation hazards. Particles falling from the eruption clouds form layers on the ground (this is pyroclastic fall or tephra). Pyroclastic density currents, which are referred to as 'flows' or 'surges' depending on particle concentration and the level turbulence, are sometimes called *glowing avalanches*. The deposits of pumice-rich pyroclastic flows can be called ignimbrites.

A pyroclastic eruption entails spitting or "fountaining" lava, where the lava will be thrown into the air along with ash, pyroclastic materials, and other volcanic byproducts. Hawaiian eruptions such as those at Kilauea can eject clots of magma suspended into gas; this is called a 'fire fountain'. The magma clots, if hot enough may coalesce upon landing to form a lava flow.

Pyroclastic deposits consist of pyroclasts which are not cemented together. Pyroclastic rocks (tuff) are pyroclastic deposits which have been lithified.

Mineralogical classification

For volcanic rocks, mineralogy is important in classifying and naming lavas. The most important criterion is the phenocryst species, followed by the groundmass mineralogy. Often, where the groundmass is aphanitic, chemical classification must be used to properly identify a volcanic rock.

Mineralogic contents - felsic versus mafic

- *felsic* rock, highest content of silicon, with predominance of quartz, alkali feldspar and/or feldspathoids: *the felsic minerals*; these rocks (e.g., granite, rhyolite) are usually light coloured, and have low density.
- *mafic* rock, lesser content of silicon relative to felsic rocks, with predominance of mafic minerals pyroxenes, olivines and calcic plagioclase; these rocks (example, basalt, gabbro) are usually dark coloured, and have a higher density than felsic rocks.
- *ultramafic* rock, lowest content of silicon, with more than 90% of mafic minerals (e.g., dunite).

For intrusive, plutonic and usually phaneritic igneous rocks where all minerals are visible at least via microscope, the mineralogy is used to classify the rock. This usually occurs on ternary diagrams, where the relative proportions of three minerals are used to classify the rock.

The following table is a simple subdivision of igneous rocks according both to their composition and mode of occurrence.

	Composition			
Mode of occurrence	Felsic	Intermediate	Mafic	Ultramafic
Intrusive	Granite	Diorite	Gabbro	Peridotite
Extrusive	Rhyolite	Andesite	Basalt	Komatiite

	Essential rock forming silicates			
	Felsic	Intermediate	Mafic	Ultramafic
Coarse Grained	Granite	Diorite	Gabbro	Peridotite
Medium Grained			Diabase	
Fine Grained	Rhyolite	Andesite	Basalt	Komatiite

Example of classification

Granite is an igneous intrusive rock (crystallized at depth), with felsic composition (rich in silica and predominately quartz plus potassium-rich feldspar plus sodium-rich plagioclase) and phaneritic, subeuhedral texture (minerals are visible to the unaided eye and commonly some of them retain original crystallographic shapes).

Magma origination

The Earth's crust averages about 35 kilometers thick under the continents, but averages only some 7-10 kilometers beneath the oceans. The continental crust is composed primarily of sedimentary rocks resting on crystalline *basement* formed of a great variety of metamorphic and igneous rocks including granulite and granite. Oceanic crust is composed primarily of basalt and gabbro. Both continental and oceanic crust rest on peridotite of the mantle.

Rocks may melt in response to a decrease in pressure, to a change in composition such as an addition of water, to an increase in temperature, or to a combination of these processes.

Other mechanisms, such as melting from impact of a meteorite, are less important today, but impacts during accretion of the Earth led to extensive melting, and the outer several hundred kilometers of our early Earth probably was an ocean of magma. Impacts of large meteorites in last few hundred million years have been proposed as one mechanism responsible for the extensive basalt magmatism of several large igneous provinces.

Decompression

Decompression melting occurs because of a decrease in pressure. The solidus temperatures of most rocks (the temperatures below which they are completely solid) increase with increasing pressure in the absence of water. Peridotite at depth in the Earth's mantle may be hotter than its solidus temperature at some shallower level. If such rock rises during the convection of solid mantle, it will cool slightly as it expands in an adiabatic process, but the cooling is only about 0.3°C per kilometer. Experimental studies of appropriate peridotite samples document that the solidus temperatures increase by 3°C to 4°C per kilometer. If the rock rises far enough, it will begin to melt. Melt droplets can coalesce into larger volumes and be intruded upwards. This process of melting from upward movement of solid mantle is critical in the evolution of Earth.

Decompression melting creates the ocean crust at mid-ocean ridges. Decompression melting caused by the rise of mantle plumes is responsible for creating ocean islands like the Hawaiian islands. Plume-related decompression melting also is the most common explanation for flood basalts and oceanic plateaus (two types of large igneous provinces), although other causes such as melting related to meteorite impact have been proposed for some of these huge volumes of igneous rock.

Effects of water and carbon dioxide

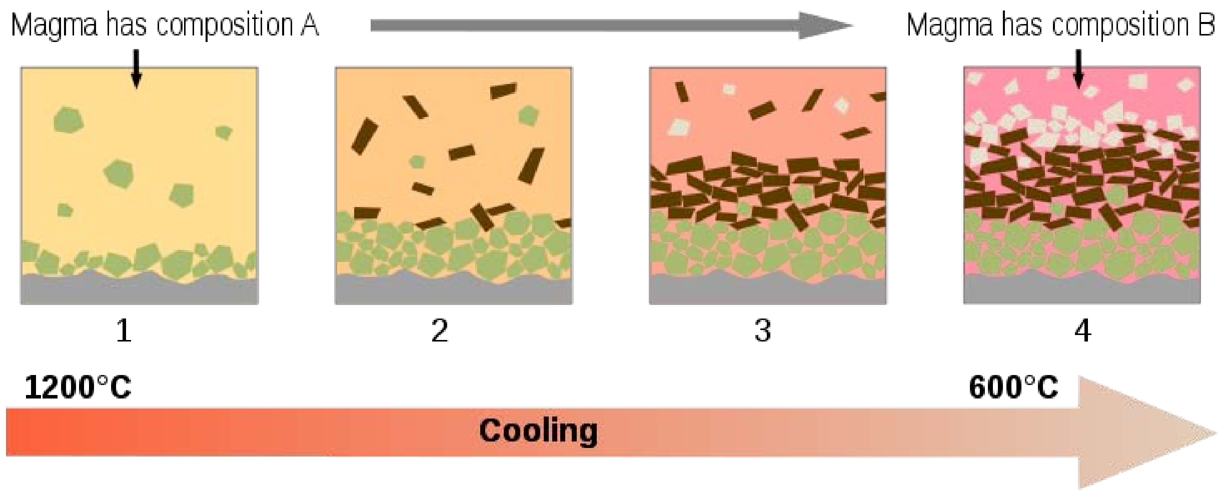
The change of rock composition most responsible for creation of magma is the addition of water. Water lowers the solidus temperature of rocks at a given pressure. For example, at a depth of about 100 kilometers, peridotite begins to melt near 800°C in the presence of excess water, but near or above about 1500°C in the absence of water. Water is driven out of the oceanic lithosphere in subduction zones, and it causes melting in the overlying mantle. Hydrous magmas of basalt and andesite composition are produced directly and indirectly as results of dehydration during the subduction process. Such magmas and those derived from them build up island arcs such as those in the Pacific ring of fire. These magmas form rocks of the calc-alkaline series, an important part of continental crust.

The addition of carbon dioxide is relatively a much less important cause of magma formation than addition of water, but genesis of some silica-undersaturated magmas has been attributed to the dominance of carbon dioxide over water in their mantle source regions. In the presence of carbon dioxide, experiments document that the peridotite solidus temperature decreases by about 200°C in a narrow pressure interval at pressures corresponding to a depth of about 70 km. At greater depths, carbon dioxide can have more effect: at depths to about 200 km, the temperatures of initial melting of a carbonated peridotite composition were determined to be 450°C to 600°C lower than for the same composition with no carbon dioxide. Magmas of rock types such as nephelinite, carbonatite, and kimberlite are among those that may be generated following an influx of carbon dioxide into mantle at depths greater than about 70 km.

Temperature increase

Increase of temperature is the most typical mechanism for formation of magma within continental crust. Such temperature increases can occur because of the upward intrusion of magma from the mantle. Temperatures can also exceed the solidus of a crustal rock in continental crust thickened by compression at a plate boundary. The plate boundary between the Indian and Asian continental masses provides a well-studied example, as the Tibetan Plateau just north of the boundary has crust about 80 kilometers thick, roughly twice the thickness of normal continental crust. Studies of electrical resistivity deduced from magnetotelluric data have detected a layer that appears to contain silicate melt and that stretches for at least 1000 kilometers within the middle crust along the southern margin of the Tibetan Plateau. Granite and rhyolite are types of igneous rock commonly interpreted as products of melting of continental crust because of increases of temperature. Temperature increases also may contribute to the melting of lithosphere dragged down in a subduction zone.

Magma evolution



Schematic diagrams showing the principles behind fractional crystallisation in a magma. While cooling, the magma evolves in composition because different minerals crystallize from the melt. **1**: olivine crystallizes; **2**: olivine and pyroxene crystallize; **3**: pyroxene and plagioclase crystallize; **4**: plagioclase crystallizes. At the bottom of the magma reservoir, a cumulate rock forms.

Most magmas only entirely melt for small parts of their histories. More typically, they are mixes of melt and crystals, and sometimes also of gas bubbles. Melt, crystals, and bubbles usually have different densities, and so they can separate as magmas evolve.

As magma cools, minerals typically crystallize from the melt at different temperatures (fractional crystallization). As minerals crystallize, the composition of the residual melt typically changes. If crystals separate from melt, then the residual melt will differ in composition from the parent magma. For instance, a magma of gabbroic composition can produce a residual melt of granitic composition if early formed crystals are separated from the magma. Gabbro may have a liquidus temperature near 1200°C, and derivative granite-composition melt may have a liquidus temperature as low as about 700°C. Incompatible elements are concentrated in the last residues of magma during fractional crystallization and in the first melts produced during partial melting: either process can form the magma that crystallizes to pegmatite, a rock type commonly enriched in incompatible elements. Bowen's reaction series is important for understanding the idealised sequence of fractional crystallisation of a magma.

Magma composition can be determined by processes other than partial melting and fractional crystallization. For instance, magmas commonly interact with rocks they intrude, both by melting those rocks and by reacting with them. Magmas of different compositions can mix with one another. In rare cases, melts can separate into two immiscible melts of contrasting compositions.

There are relatively few minerals that are important in the formation of common igneous rocks, because the magma from which the minerals crystallize is rich in only certain elements: silicon, oxygen, aluminium, sodium, potassium, calcium, iron, and magnesium. These are the elements which combine to form the silicate minerals, which account for over ninety percent of all igneous rocks. The chemistry of igneous rocks is expressed differently for major and minor elements and for trace elements. Contents of major and minor elements are conventionally expressed as weight percent oxides (e.g., 51% SiO_2, and 1.50% TiO_2). Abundances of trace elements are conventionally expressed as parts per million by weight (e.g., 420 ppm Ni, and 5.1 ppm Sm). The term "trace element" typically is used for elements present in most rocks at abundances less than 100 ppm or so, but some trace elements may be present in some rocks at abundances exceeding 1000 ppm. The diversity of rock compositions has been defined by a huge mass of analytical data—over 230,000 rock analyses can be accessed on the web through a site sponsored by the U. S. National Science Foundation.

Etymology

The word "igneous" is derived from the Latin *ignis*, meaning "of fire". Volcanic rocks are named after Vulcan, the Roman name for the god of fire.

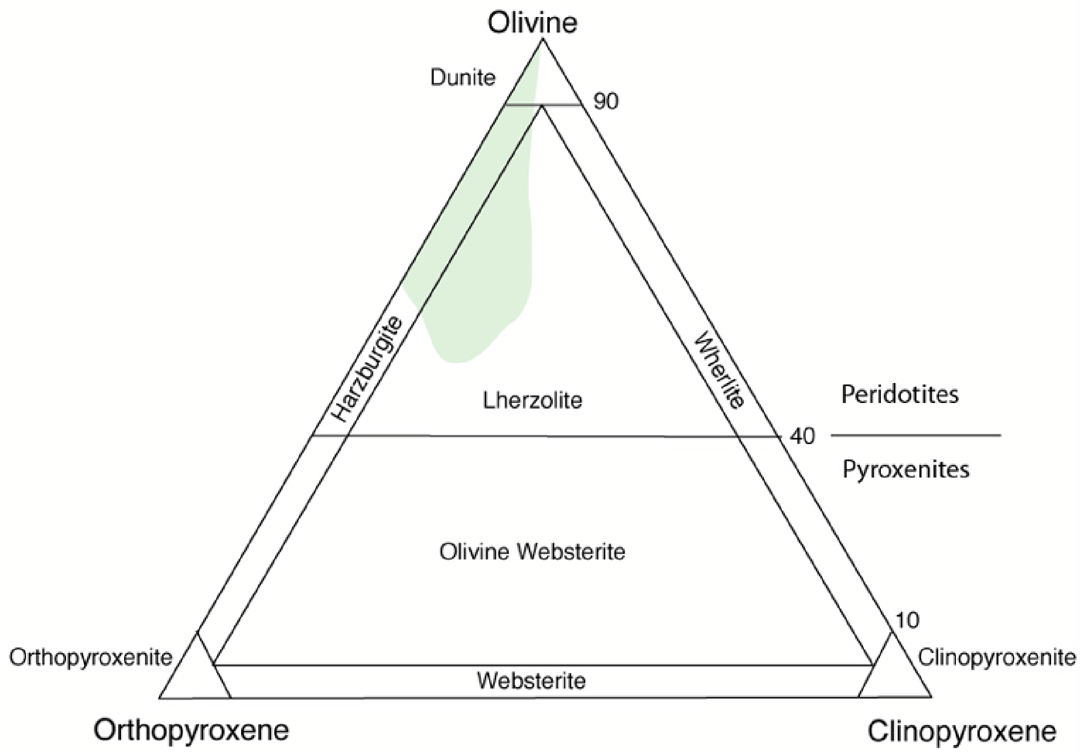
Intrusive rocks are also called plutonic rocks, named after Pluto, the Roman god of the underworld.

Ultramafic rock

Olivine in a peridotite weathering to iddingsite within a mantle xenolith

Ultramafic (also referred to as **ultrabasic**) rocks are igneous and meta-igneous rocks with very low silica content (less than 45%), generally >18% MgO, high FeO, low potassium, and are composed of usually greater than 90% mafic minerals (dark colored, high magnesium and iron content). The Earth's mantle is composed of ultramafic rocks.

Intrusive ultramafic rocks



IUGS Classification diagram for intrusive ultramafic rocks based on modal percentages of mafic minerals. Green area represents typical mantle peridotite.

Intrusive ultramafic rocks are often found in large, layered ultramafic intrusions where differentiated rock types often occur in layers. Such cumulate rock types do not represent the chemistry of the magma from which they crystallized. The ultramafic intrusives include the dunites, peridotites and pyroxenites. Other rare varieties include troctolite which has a greater percentage of calcic plagioclase. These grade into the anorthosites. Gabbro and norite often occur in the upper portions of the layered ultramafic sequences. Hornblendite and, rarely phlogopitite, are also found.

Volcanic ultramafic rocks

Volcanic ultramafic rocks are rare outside of the Archaean and are essentially restricted to the Neoproterozoic or earlier, although some boninite lavas currently erupted within back-arc basins (Manus Trough, New Guinea) verge on being ultramafic. Subvolcanic ultramafic rocks and dykes persist longer, but are also rare. Many of the lavas being produced on Io may be ultramafic, as evidenced by their temperatures which are higher than terrestrial mafic eruptions.

Examples include komatiite and picritic basalt. Komatiites can be host to ore deposits of nickel.

Ultrapotassic ultramafic rocks

Technically ultrapotassic rocks and melilitic rocks are considered a separate group, based on melting model criteria, but there are ultrapotassic and highly silica-under-saturated rocks with >18% MgO. which can be considered "ultramafic".

Ultrapotassic, ultramafic igneous rocks such as lamprophyre, lamproite and kimberlite are known to have reached the surface of the Earth. Although no modern eruptions have been observed, analogues are preserved.

Most of these rocks occur as dikes, diatremes, lopoliths or laccoliths, and very rarely, intrusions. Most kimberlite and lampproite occurrences occur as volcanic and subvolcanic diatremes and maars; lavas are virtually unknown.

Vents of Proterozoic lamproite (Argyle diamond mine), and Cenozoic lamproite (Gaussberg, Antarctica) are known, as are vents of Devonian lamprophyre (Scotland). Kimberlite pipes in Canada, Russia and South Africa have incompletely-preserved tephra and agglomerate facies.

These are generally diatreme events and as such are not lava flows although tephra and ash deposits are partially preserved. These represent low-volume volatile melts and attain their ultramafic chemistry via a different process to typical ultramafic rocks.

Carbonatites are rare high-carbonate, low-silica igneous rocks.

Metamorphic ultramafic rocks

Metamorphism of ultramafic rocks in the presence of water and/or carbon dioxide results in two main classes of metamorphic ultramafic rock; talc carbonate and serpentinite.

Talc carbonation reactions occur in ultramafic rocks at lower greenschist through to granulite facies metamorphism when the rock in question is subjected to metamorphism and the metamorphic fluid has more than 10% molar proportion of carbon dioxide.

When the metamorphic fluids in contact with the ultramafic rock have less than 10% CO_2 the metamorphic reactions favor serpentinisation reactions, resulting in chlorite-serpentine-amphibole type assemblages.

Distribution in space and time

The majority of ultramafic rocks are exposed in orogenic belts, and predominate in Archaean and Proterozoic terranes. Ultramafic magmas in the Phanerozoic are rarer, and there are very few recognised true ultramafic lavas in the Phanerozoic.

Many surface exposures of ultramafic rocks occur in ophiolite complexes where deep mantle-derived rocks have been obducted onto continental crust along and above subduction zones.

Ultramafic rocks and the regolith

Alkali-rich ultramafic rock can provide an excellent balance of nutrients to the soils that develop on them. But the more common peridotites and serpentinites have an excessively high ratio of magnesium to calcium along with deficiencies of phosphorus and potassium. There also are toxic amounts of chromium and nickel. These factors create unique vegetation. Examples are the ultramafic woodlands and ultramafic barrens of the Appalachian mountains and piedmont, the "wet maquis" of the New Caledonia rain forests, and the ultramafic forests of Mount Kinabalu and other peaks in Sabah, Malaysia. Vegetation is typically stunted, and is sometimes home to endemic species adapted to the metallic soils.

Often thick, magnesite-calcrete caprock, clayey laterite and duricrust forms over ultramafic rocks in tropical and subtropical environments. Particular floral assemblages associated with highly nickeliferous ultramafic rocks are indicative tools for mineral exploration.

Weathered ultramafic rocks may form lateritic nickel ore deposits

Chapter- 6

Shear Zone

A **shear zone** is a very important structural discontinuity surface in the Earth's crust and upper mantle. It forms as a response to inhomogeneous deformation partitioning strain into planar or curviplanar high-strain zones. Intervening (crustal) blocks stay relatively unaffected by the deformation. Due to the shearing motion of the surrounding more rigid medium, a rotational, non co-axial component can be induced in the shear zone. Because the discontinuity surface usually passes through a wide depth-range, a great variety of different rock types with their characteristic structures are produced.

General introduction

A shear zone is a zone of strong deformation (with a high strain rate) surrounded by rocks with a lower state of finite strain. It is characterised by a length to width ratio of more than 5:1.

Shear zones form a continuum of geological structures, ranging from *brittle shear zones* (or faults) via *brittle–ductile shear zones* (or *semibrittle shear zones*), *ductile–brittle* to *ductile shear zones*. In brittle shear zones, the deformation is concentrated in a narrow fracture surface separating the wall rocks, whereas in a ductile shear zone the deformation is spread out through a wider zone, the deformation state varying continuously from wall to wall. Between these end-members, there are intermediate types of brittle–ductile (semibrittle) and ductile–brittle shear zones that can combine these geometric features in different proportions.

This continuum found in the structural geometries of shear zones reflects the different deformation mechanisms reigning in the crust, i.e. the changeover from brittle (fracturing) at or near the surface to ductile (flow) deformation with increasing depth. By passing through the *brittle–semibrittle transition* the ductile response to deformation is starting to set in. This transition is not tied to a specific depth, but rather occurs over a certain depth range - the so-called **alternating zone**, where brittle fracturing and plastic flow coexist. The main reason for this is found in the usually heteromineral composition of rocks, with different minerals showing different responses to applied stresses (for instance, under stress quartz reacts plastically long before feldspars do). Thus differences in lithology, grain size, and preexisting fabrics determine a different rheological response. Yet other, purely physical factors, influence the changeover depth as well, including:

- geothermal gradient, i.e. ambient temperature.
- confinement pressure and fluid pressure.
- bulk strain rate.
- stress field orientation.

In Scholz's model for a quartzo-feldspathic crust (with a geotherm taken from Southern California), the brittle–semibrittle transition starts at about 11 km depth with an ambient temperature of 300°C. The underlying alternating zone then extends to roughly 16 km depth with a temperature of about 360°C. Below approximately 16 km depth, only ductile shear zones are found.

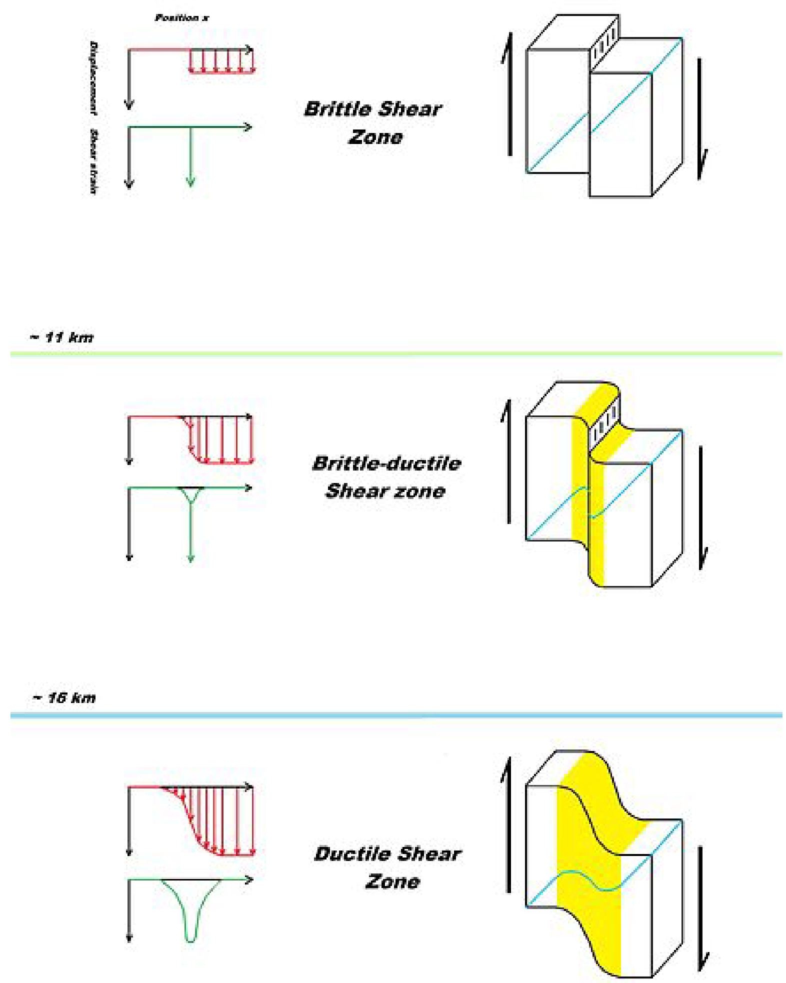


Diagram showing the major different types of shear zones. Displacement, shear strain, and depth distribution are also indicated.

The *seismogenic zone*, in which earthquakes nucleate, is tied to the brittle domain, the schizosphere. Below an intervening alternating zone, there is the plastosphere. In the seismogenic layer, which occurs below an *upper stability transition* related to an upper seismicity cutoff (situated usually at about 4–5 km depth), true cataclasites start to appear. The seismogenic layer then yields to the alternating zone at 11 km depth. Yet big earthquakes can rupture both up to the surface and well into the alternating zone, sometimes even into the plastosphere.

Rocks produced in shear zones

The deformations in shear zones are responsible for the development of characteristic fabrics and mineral assemblages reflecting the reigning pressure–temperature (pT) conditions, flow type, movement sense, and deformation history. Shear zones are therefore very important structures for unravelling the history of a specific terrane.

Starting at the Earth's surface, the following rock types are usually encountered in a shear zone:

- uncohesive fault rocks. Examples being fault gouge, fault breccia, and foliated gouge.
- cohesive fault rocks like crush breccias and cataclasites (protocataclasite, cataclasite, and ultracataclasite).
- glassy pseudotachylites.

Both fault gouge and cataclasites are due to abrasive wear on brittle, seismogenic faults.

- foliated mylonites (phyllonites).
- striped gneiss.

Mylonites start to occur with the onset of semibrittle behaviour in the alternating zone characterised by adhesive wear. Pseudotachylites can still be encountered here. By passing into greenschist facies conditions, the pseudotachylites disappear and only different types of mylonites persist. Striped gneisses are high-grade mylonites and occur at the very bottom of ductile shear zones.

Sense of shear

The sense of shear in a shear zone (dextral, sinistral, reverse or normal) can be deduced by macroscopic structures and by a plethora of microtectonic indicators.

Indicators

The main macroscopic indicators are striations (slickensides), slickenfibers, and stretching– or mineral lineations. They indicate the direction of movement. With the aid of offset markers such as displaced layering and dykes, or the deflection (bending) of layering/foliation into a shear zone, one can additionally determine the sense of shear.

En echelon tension gash arrays (or extensional veins), characteristic of ductile-brittle shear zones, and sheath folds can also be valuable macroscopic shear-sense indicators.

Microscopic indicators consist of the following structures:

- asymmetric folds.
- foliations.
- imbrications.
- Crystallographic preferred orientation (CPO).
- mantled and winged porphyroclasts. Well-known examples are theta (Θ)-objects and phi (Φ)-porphyroclasts, as well as sigma (σ)- and delta (δ)-winged objects.
- mica-fish (foliation fish).
- pressure shadows
- pull-aparts.
- quarter structures.
- shear band cleavages.
- step-over sites.

Width of shear zones and resulting displacements

The width of individual shear zones stretches from the grain scale to the kilometer scale. Crustal-scale shear zones (megashears) can become 10 km wide and consequently show very large displacements from tens to hundreds of kilometers.

Brittle shear zones (faults) usually widen with depth and with an increase in displacements.

Strain softening and ductility

Because shear zones are characterised by the localisation of strain, some form of *strain softening* must occur, in order for the affected host material to deform more plastically. The softening can be brought about by the following phenomena:

- grain-size reductions.
- geometric softening.
- reaction softening.
- fluid-related softening.

Furthermore for a material to become more ductile (quasi-plastic) and undergo continuous deformation (flow) without fracturing, the following deformation mechanisms (on a grain scale) have to be taken into account:

- diffusion creep (various types).
- dislocation creep (various types).
- dynamic recrystallization
- pressure solution processes.
- grain-boundary sliding (superplasticity) and grain-boundary area reduction.

Occurrence and examples of shear zones

Due to their deep penetration, shear zones are found in all metamorphic facies. Brittle shear zones are more or less ubiquitous in the upper crust. Ductile shear zones start at greenschist facies conditions and are therefore bound to metamorphic terranes.

Shear zones can occur in the following geotectonic settings:

- transcurrent setting – steep to vertical:
    - strike-slip zones.
    - transform faults.

- compressive setting — low-angle
    - recumbent fold nappes (at the base of).
    - subduction zones.
    - thrust sheets (at the base of).

- extensional setting — low-angle
    - metamorphic core complex detachments.

Shear zones are dependent neither on rock type nor on geological age. Most often they are not isolated in their occurrence, but commonly form fractal-scaled, linked up, *anastomosing networks* which reflect in their arrangement the underlying dominant sense of movement of the terrane at that time.

Some good examples of shear zones of the strike-slip type are the South Armorican Shear Zone and the North Armorican Shear Zone in Brittany, the North Anatolian Fault Zone in Turkey, and the Dead Sea Fault in Israel. Shear zones of the transform type are the San Andreas Fault in California, and the Alpine Fault in New Zealand. A shear zone of the thrust type is the Moine Thrust in northwestern Scotland. An example for the subduction zone setting is the Japan Median Tectonic Line. Detachment fault related shear zones can be found in southeastern California, e.g. the Whipple Mountain Detachment Fault. An example of a huge anastomosing shear-zone is the Borborema Shear Zone in Brazil.

Importance

The importance of shear zones lies in the fact that they are major zones of weakness in the Earth's crust, sometimes extending into the upper mantle. They can be very long-lived features and commonly show evidence of several overprinting stages of activity. Material can be transported upwards or downwards in them, the most important one being water circulating dissolved ions. This can bring about metasomatism in the host rocks and even re-fertilise mantle material.

Shear zones can host economically viable mineralizations, examples being important gold deposits in Precambrian terranes.