Introduction to Structural Geology

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Introduction

The Place of Structural Geology in Sciences

Science is the search for knowledge about the Universe, its origin, its evolution, and how it works. Geology, one of the core science disciplines with physics, chemistry, and biology, is the search for knowledge about the Earth, how it formed, evolved, and how it works. Geology is often presented in the broader context of Geosciences; a grouping of disciplines specifically looking for knowledge about the interaction between Earth processes, environments and societies.

Structural Geology, Tectonics and Geodynamics form a coherent and interdependent ensemble of sub-disciplines, the aim of which is the search for knowledge about how minerals, rocks and rock formations, and Earth systems (i.e., crust, lithosphere, asthenosphere ...) deform and through which processes.
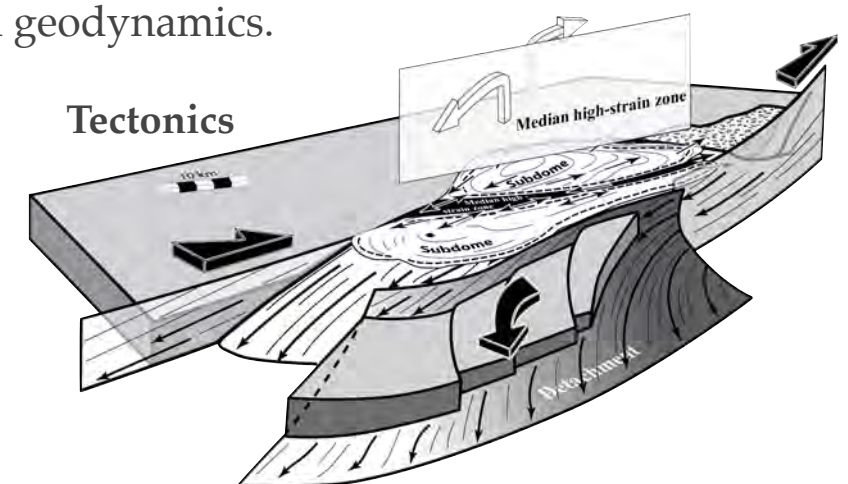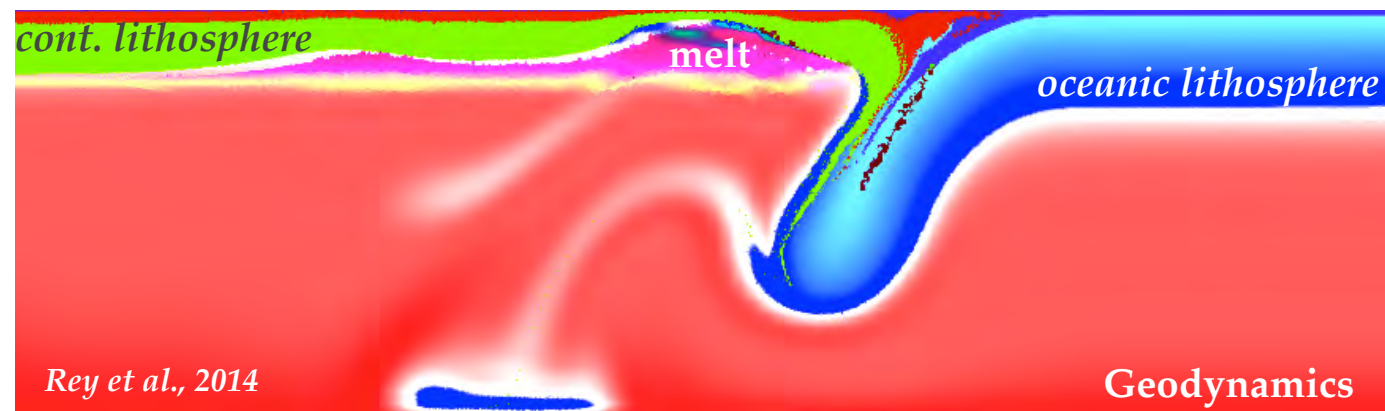
Structural Geology In Geosciences

Structural Geology aims to characterise deformation structures (*geometry*), to characterize flow paths followed by particles during deformation (*kinematics*), and to infer the direction and magnitude of the forces involved in driving deformation (*dynamics*). As a field-based discipline, structural geology operates at scales ranging from 100 microns to 100 meters (i.e. grain to outcrop).

Tectonics aims at unraveling the geological context in which deformation occurs. It involves the integration of structural geology data in maps, cross-sections and 3D block diagrams, as well as data from other Geoscience disciplines including sedimentology, petrology, geochronology, geochemistry and geophysics. Tectonics operates at scales ranging from 100 m to 1000 km, and focus on processes such as continental rifting and basin formation, subduction, collisional processes and mountain building processes etc.

Geodynamics focus on the forces that drive mantle convection, plate motion and deformation of Earth's material. Geodynamics is concerned with deep mantle processes such as mantle convection, cold drips, hot plumes and their links to plate motion, including dynamic plate subsidence and uplift, and plate tectonic processes. Geodynamics involves working at scales > 100 km. Numerical modeling is at the core of modern geodynamics.

Structural geology

cont. lithosphere

melt

oceanic lithosphere

Rey et al., 2014

Geodynamics

Tectonics

Median high-strain zone

The Scientific Approach

Like all scientists, structural geologists follow research strategies that call upon concepts such as: fact, hypothesis, model, theory, and law. A good understanding of these terms is essential for all scientists.

A **fact** is a bit of truth. For a structural geologist a "fact" could be the dip direction of a bedding plane. Having measured that north is to the left on the top image, then it is a fact that the dip direction of the bedding is to the south.

An **hypothesis** is an assumed fact. It is a short statement one makes to go further into a reasoning. For instance, assuming that the bottom photo is parallel to the stretching lineation (hypothesis), then one can infer that the sense of shear deduced from the "tiling" of K-feldspars in this orthogneiss is top to the right.

A **model** is a self-consistent framework providing a coherent explanation for the observed facts. A model combines as many facts as possible and as little hypotheses as possible. A good model enables predictions to be made. A model is proven wrong if key predictions are not verified. It can be iteratively strengthened via minor modifications to account for both the facts and the predictions.

<<<<--- Stretching lineation --- >>>

A **theory** is a very robust model which accounts for a large number of independent facts; and whose numerous predictions have been verified over a long period of time. *Evolution* and *Plate Tectonics* are two scientific theories, *Creationism* is not.

A **law** is a simple, fundamental concept that is always verified by experiments and that underpins our understanding of the world. For instance, the law of gravity and the laws of thermodynamics underpin our understanding of physics and chemistry.

Truth is not a scientific concept.

Scientific Methods: Francis Bacon (1561-1626) vs Charles Darwin (1809-1882)

The **Baconian method**: This method, sometimes called the "*induction approach*", is based on the collection of facts without regard to a particular model. Eventually, the growing dataset self organizes into a model; this is the "*zero hypothesis*" approach. Isaac Newton, a fervent adept of Francis Bacon's induction approach, once said: "*I don't make hypotheses*". This method is at the core of modern data mining.

The **Darwinian method**: A model is built from of a set of facts. The model is used to make verifiable predictions. The predictions are verified by the acquisition of new data or via experimentation, if necessary the model is modified to account for the new data or the new observation. This iterative process eventually leads to a robust model.

There is no need to oppose the Baconian and the Darwinian methods. Both have earned their place in Science.

The **Reality**...: Scientific progress is almost never the result of strictly rational scientific methods. Intuition, non-rational and non-logical thinking make science fun, exciting and surprising. Einstein once said: "*The only real valuable thing is intuition.*" ... and also that ... "*Imagination is more important than knowledge.*"
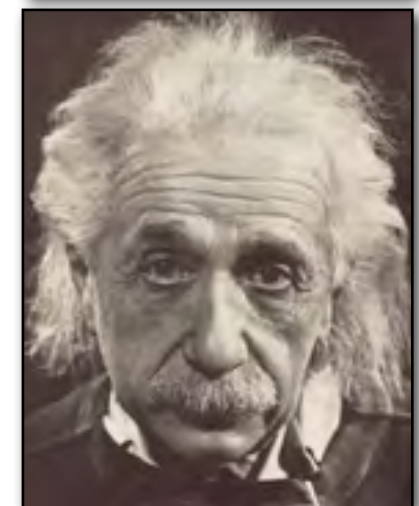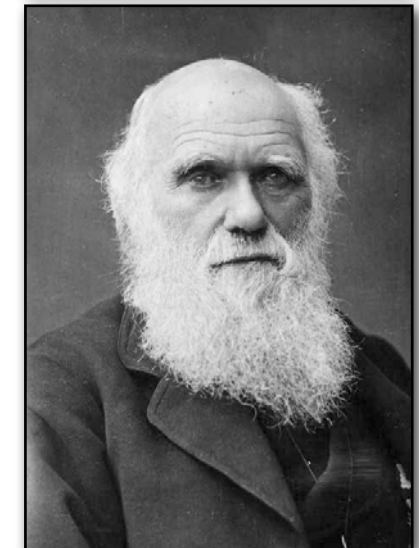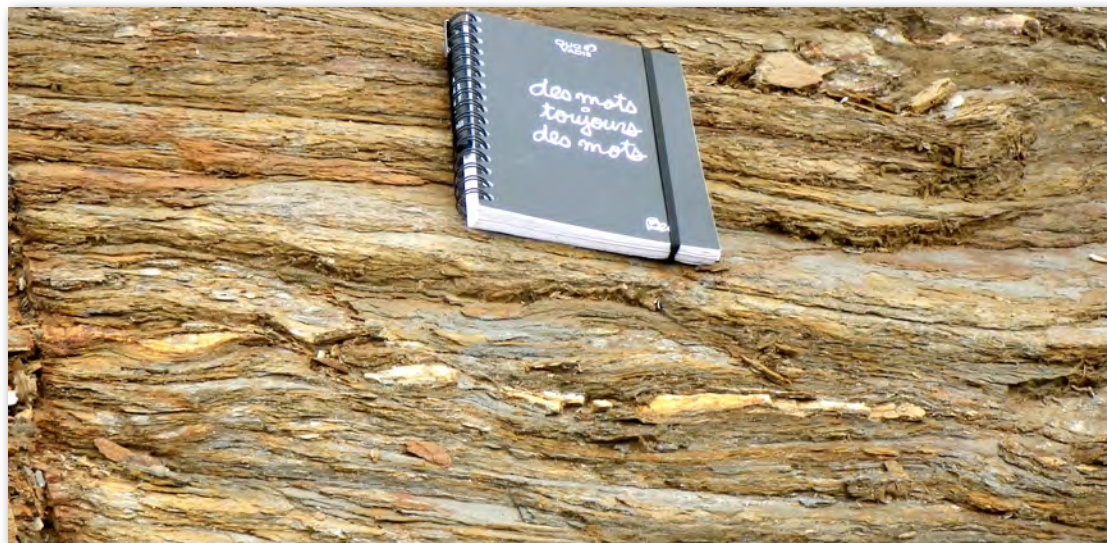
Workflow Of Structural Geology & Tectonics

Structural Geology and Tectonics combine two aspects:

1/ Description and analysis of 3D structures and microstructures (Structural Geology *sensu stricto*). Structural geologists are concerned with features resulting from deformation. These include fractures, faults, folds, boudins, shear zones, cleavages (also knows as schistosities), foliations, lineations and crystallographic fabrics (e.g. crystallographic preferred orientation).

From the analysis of these structures, they aim to understand **finite strain** (i.e., the ultimate product of long, sometimes polyphased deformation histories), and **incremental strain** (i.e., the small increments of deformation, the accumulation of which leads to the finite strain).

Structural geologists are interested in understanding "**strain fields**" by mapping deformation features such as foliations and stretching lineations that tell us the orientation of the principal shortening direction and principal lengthening direction respectively.

In the case of faults and shear zones, they are interested in understanding their **kinematics** (i.e., the relative sense of motion of the blocks on either side of faults or shear zones) and the magnitude of the displacement involved.

They are interested in inferring the direction of maximum and minimum stress directions from small deformation features such as centimeter-scale extensional fractures and associated stylolitic joints.
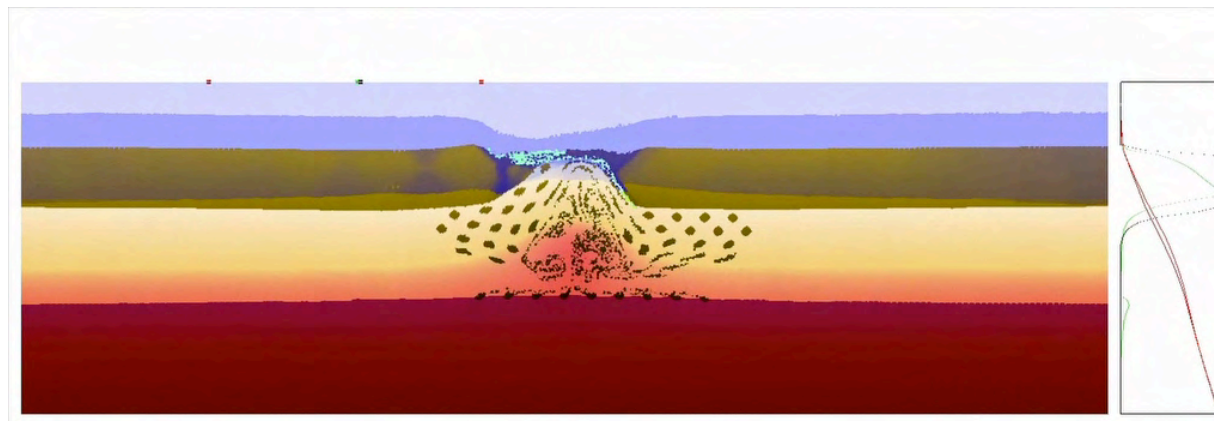
Basaltic dike intrusive in shales, (South France). A dike is an extensional fracture that was filled with lava at the time of opening. The direction of open-up is parallel to the direction of minimum stress (Photo credit: P. Rey).

2/ Design of tectonic models (Tectonics).

The purpose of these models is to explain the deformation history that led to the observed 3D strain fields. Tectonic models incorporate a broad range of data from other disciplines. No matter how tectonicists design these models - following hours of pure rational thinking or via flashes of insight after a heavy night - tectonic models should always be:

- **Physically valid**: They must obey the laws of physics. It sounds trivial but this requires serious computational modelling.

- **Testable**: They must provide testable predictions (structural, sedimentological, petrological, geochemical, geophysical ...) that can be verified.

- **Robust**: They must explain a large number of unrelated facts.

- **Lean**: Hypotheses should be kept to a minimum compared to the number of facts models explain.

The computer model predicts a double-dome structure for migmatite-cored gneiss dome. The Entia dome in central Australia has a double-dome internal structure.

Movie 1.1 A predictive model of a gneiss dome (Metamorphic core complex)

Computational tectonic model of a gneiss dome (Rey, Teyssier, Kruckenberg and Whitney, Geology, 2011).
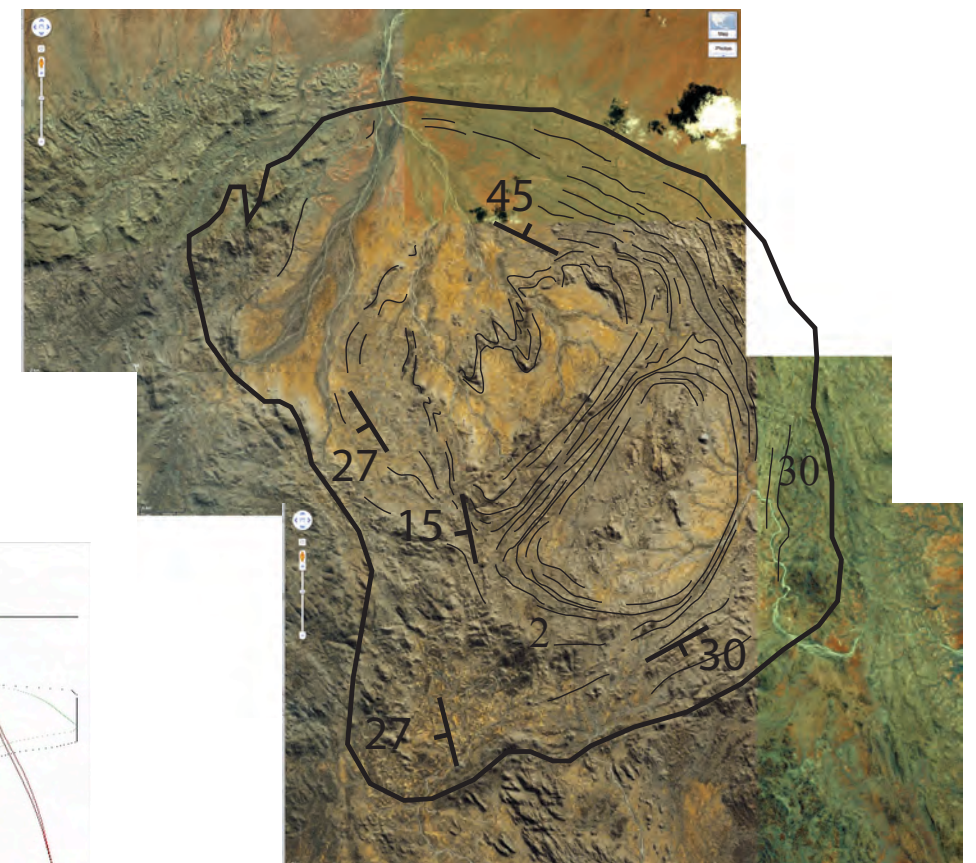
Methods Of Structural Geology & Tectonics

Data acquisition: Field Geology

The world of structural geology is by nature three dimensional (3D). To unravel 3D architecture, structural geologists process a lot of structural measurements.

In the field, structural geologists measure the strike-dip-dip direction of planar features (bedding, cleavage, fault, fold axial surface ...), and plunge plunge direction of linear features (fold axes, intersecting lineations...).

They gather information about the orientation and, when possible, the magnitude of strain and stress, and they determine the relative sense of displacement across brittle faults and ductile shear zones.

Based on cross-cutting relationships, they determine the sequence of geological and deformational events, and they characterise incremental strain and finite (i.e. accumulated) strain.

On aerial photographs and satellite images, structures too large to be seen in the field become obvious (sometimes). Ultimately, 3D structures must be consistent across all scales.
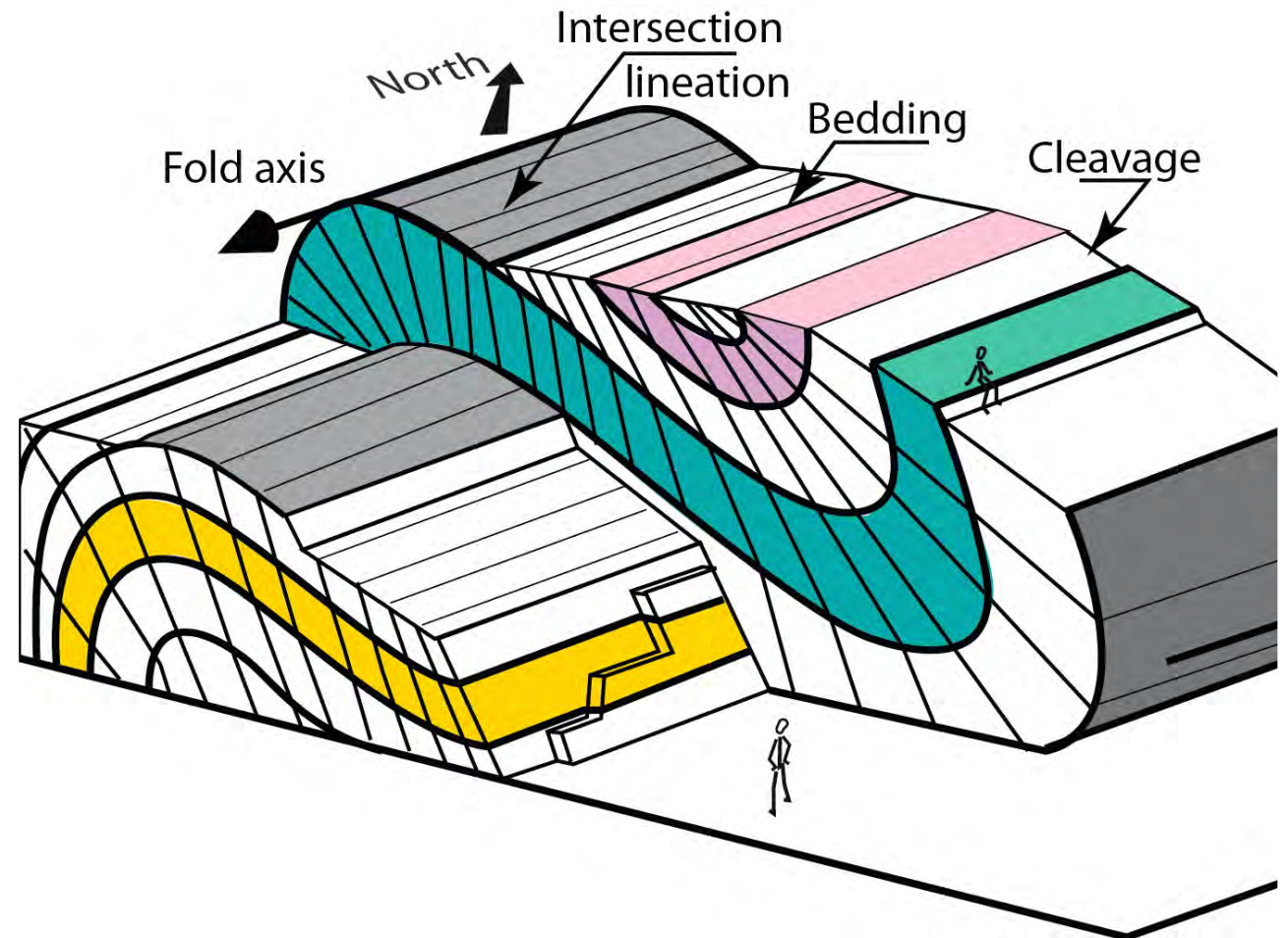
Planar surfaces are oriented in space by their strike (azimuth of a horizontal line on the planar surface), dip (angle from horizontal down to the plane) and dip direction (geographic direction toward which the plane is sloping). Folds are oriented by the strike-dip-dip direction of their axial plane, and the plunge and plunge direction of their axis (line running along their hinge).
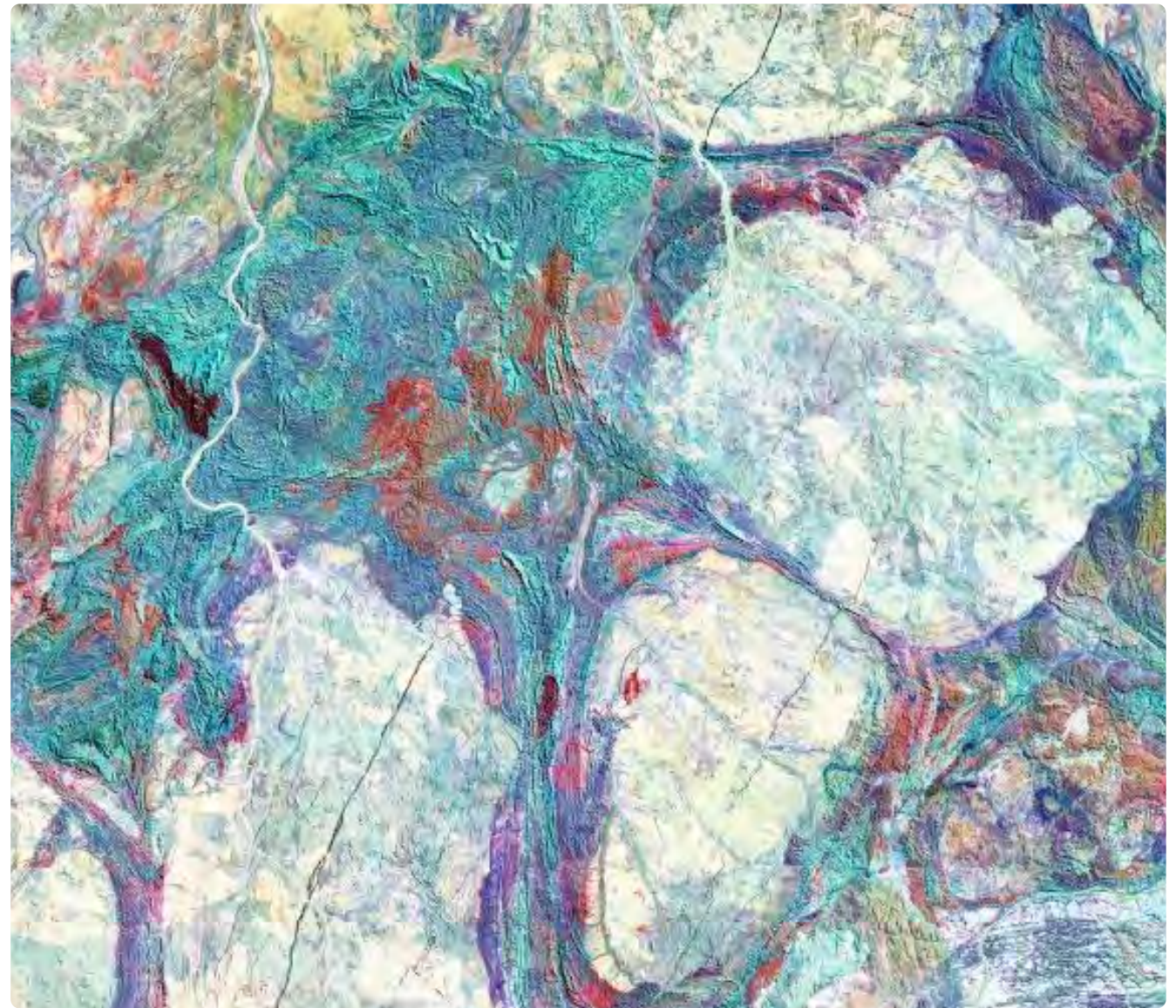
Data acquisition: Geophysics

When the surface geology is buried underneath a regolith (in-situ blanket of weathered rocks) or vegetation cover, geophysical images such as gravity anomaly maps, radiometric maps, aeromagnetic maps etc can reveal a lot of information on the nature of the sub-surface geology and deeper structures.

Remote sensing techniques (airborne and satellite) are expanding very fast. Multispectral (e.g., ASTER satellite) and hyperspectral imaging (airborne) allow geologists to have access to high-resolution images over a range of electromagnetic wavelengths covering visible (400-700 nanometer), infrared to near infrared (NIR: 700-1000 nanometer), shortwave infrared (SWIR: 1000-2500 nanometer) and far infrared (FIT: >2500 nanometer). These images allow for the mapping of the distribution of minerals at the Earth surface.

(http://portal.auscope.org/portal/gmap.html).

Gallery 1.3 Structural data from geophysics

Landsat image of the east Pilbara (WA). The pale rounded features are granitic domes. In blue green are greenstone belts (continental flood basalts). The straight lines running SSW-NNE are Proterozoic dikes.

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Data acquisition: Geophysics (cont.)

Information about deep geology is gathered directly via drilling and indirectly via geophysical methods such as gravimetry and seismic surveys amongst others. Gravimetry gives information about the distribution of masses below the surface, whereas refraction and reflection seismology reveal variation in rocks' elastic properties.

Seismic reflection is at the core of oil and gas exploration. Elastic waves, artificially induced at the Earth's surface via explosive or vibration trucks, propagate through rock formations reflecting and refracting at acoustic impedance contrasts (i.e. contrast in the product of *velocity* x *density*). The processing of waves traveling back to the surface can reveal in minute details the underground geology.

Visit the Virtual Seismic Atlas:

see-atlas.leeds.ac.uk:8080

Gallery 1.4 Structural data from geophysics
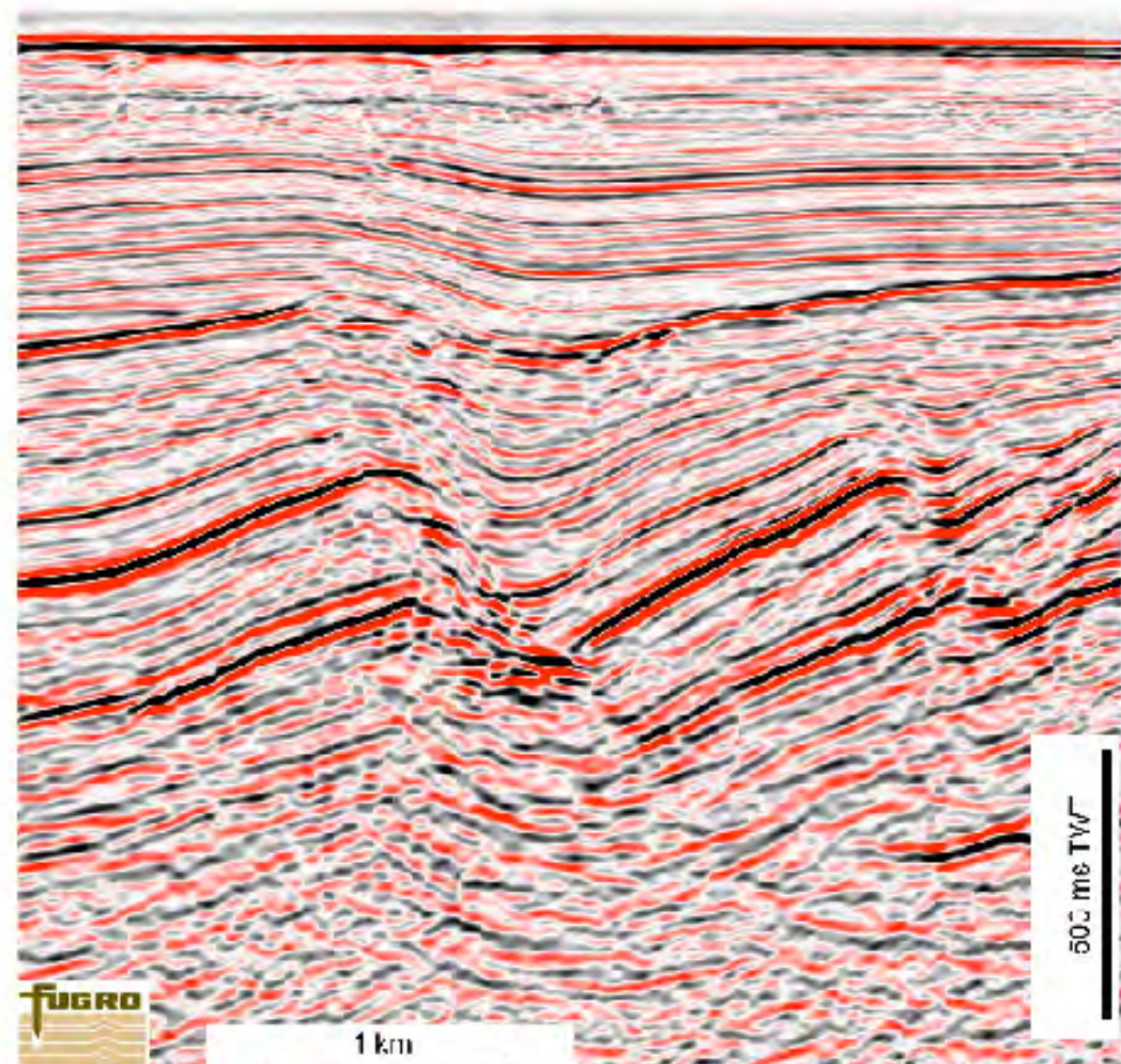
Virtual Seismic Atlas Author: Rob Butler. Normal fault in the Inner Moray Firth using Fugro 2D data.

Data synthesis and analysis :

Structural data are summarized in various documents:

1/ Geological maps and cross sections show the distribution of rock formations and structures (e,g, faults and folds) at the Earth's surface. A geological map is a very powerful document, as one can reconstruct the 3D subsurface geology.

2/ Tectonic maps and structural profiles show the distribution of deformation-related structures such as faults and shear zones, foliations and lineations, distribution of fold axes, strain magnitude etc.

3/ Structural sketch-maps are simplified geological maps in which rock formations are grouped in packages separated by unconformities, each group sharing the same tectonic history.

4/ Block diagrams, tectonograms: Illustrative 3D representation of structures.
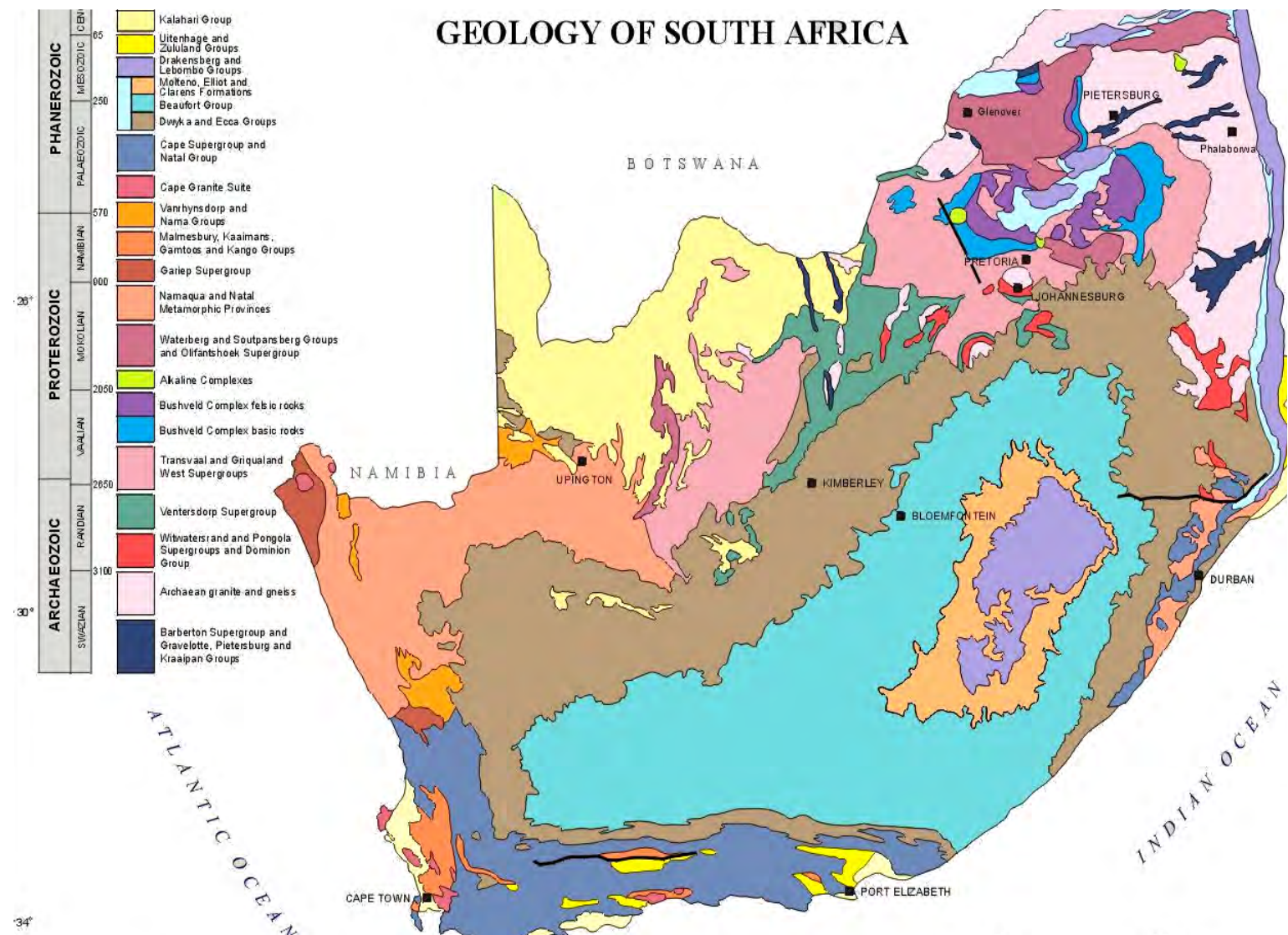
*Geological map of South Africa.*

Model generation and validation :

Geological maps and cross-sections are models. They combine data (measurements and observations made directly in the field, and from drill cores) and hypotheses. Indeed, in places where no rock is exposed geologists must *hypothesize* to fill-in observation gaps.

To validate maps and cross-sections geologists use the concept of surface conservation. This concept is based on the premise that a rock layer maintains its surface area during deformation. This premise holds true as long as rock formations deform with little in-bed mass transfer (via dissolution for instance). A cross-section is said to be balanced (i.e. geometrically consistent) when, upon restoration (i.e. retro-deformation), lithological interfaces recover their original horizontal position, and when there is no overlap or gap.

A 4D tectonic model (3D plus time), often presented in the form of a time sequence of block diagrams, must also be thermally and mechanically consistent. Verifying both aspects requires computational modelling, which in addition to its capacity of validation presents the additional advantage to be predictive.

Gallery 1.6 Model Validation
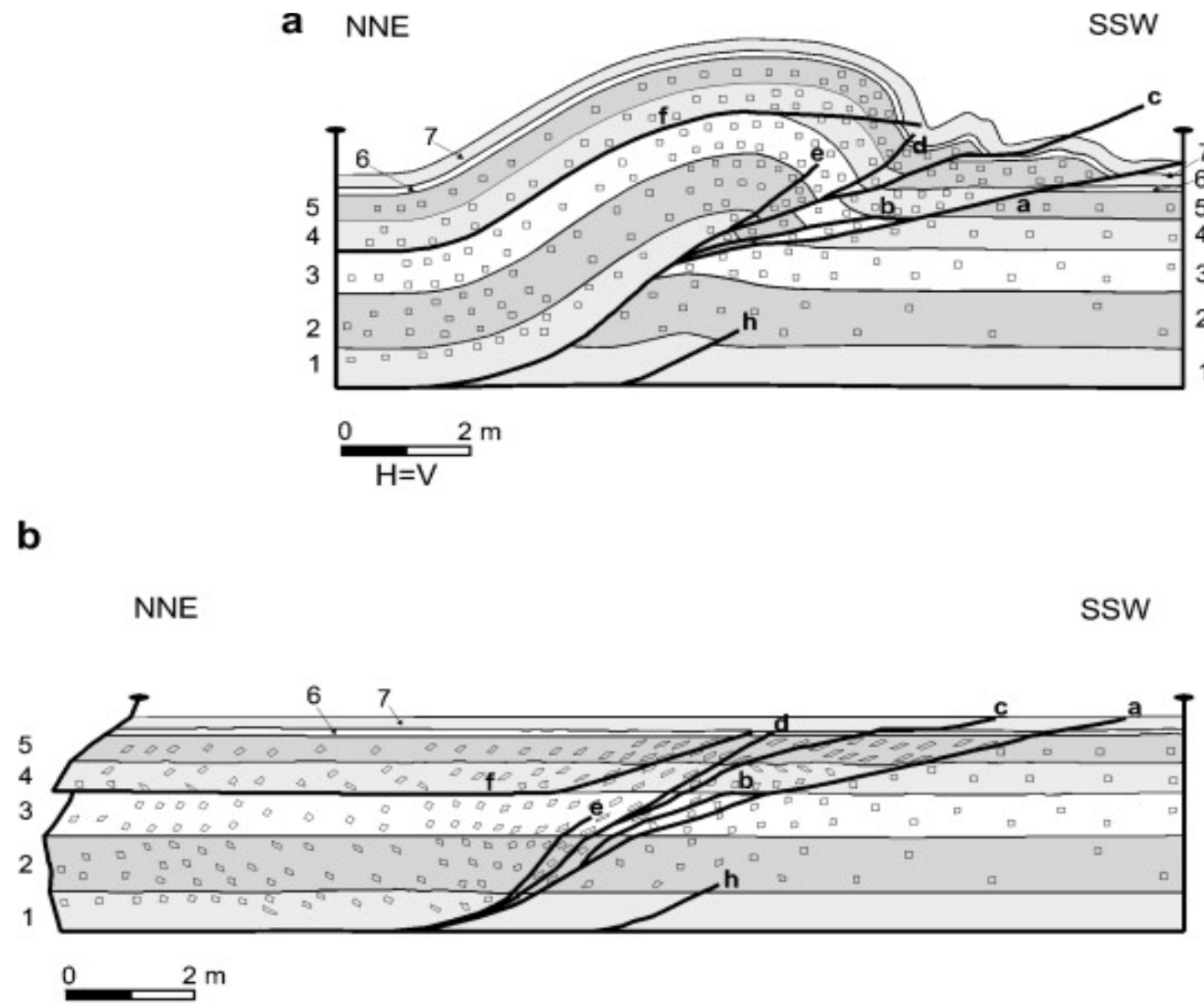
Example of cross-section restoration in a fault-propagation fold (Masini et al., Journal of Structural Geology, 2010).

Who Needs Structural Geology ?

•Structural geology is at the core of hydrocarbon and mineral exploration, as structures control the migration, trapping and escape of hydrocarbon and mineralized fluids. Structural geology is the first stage of any regional geophysical and geochemical surveys aiming at identifying new mineralized provinces. It is also critical for the interpretation of geophysical, geochemical, and geochronological data. At the mine camp scale, structural geology guides the mining process.

•Structural geology is at the core of geotechnical site assessment before the building of bridges, dams, tunnels, nuclear reactors, waste disposals etc.

• Because of the obvious relationship between faults and earthquakes, structural geology is at the core of earthquake mitigation and earthquakes seismology.

•Structural geology is central to studies on past and present mountain belts and sedimentary basins. No geological, geochemical or geophysical study can be done without the input of structural geology.

What makes a good structural geologist?

•The ability to think in 3D and to solve large-scale 4D puzzles.

•The ability to interact with a large range of geoscientists over a wide range of geological and environmental problems.

•The ability to link field studies to computational modelling.

Aims & Objectives

The objective of this eBook is to provide a robust and enjoyable initiation to *Structural Geology.*

Its readers will develop a good understanding of structures and microstructures that result from the deformation of the Earth's lithosphere. Readers will gather enough insight and knowledge to learn independently from advanced Structural Geology textbooks.

This eBook covers the basics about faults, folds, shear zones, boudins and their related fabrics. We will learn to use fractures and faults to conduct paleostress analysis, a fundamental constraint for hydrocarbon and mineral exploration, as well as risk assessment in seismically active regions. We will learn to use folds, shear zones and fabrics to perform strain analysis in order to understand strain regime and strain magnitude.

Finally, this eBook will motivate its readers to go into the field and understand geology from a structural geology perspective.

Question 1 of 5

In Science, a *Theory* is:

   ○   **A.** An hypothesis still to be tested.

   ○   **B.** An assumption based on facts.

   ✓   **C.** A robust model.

   ○   **D.** A proven hypothesis.

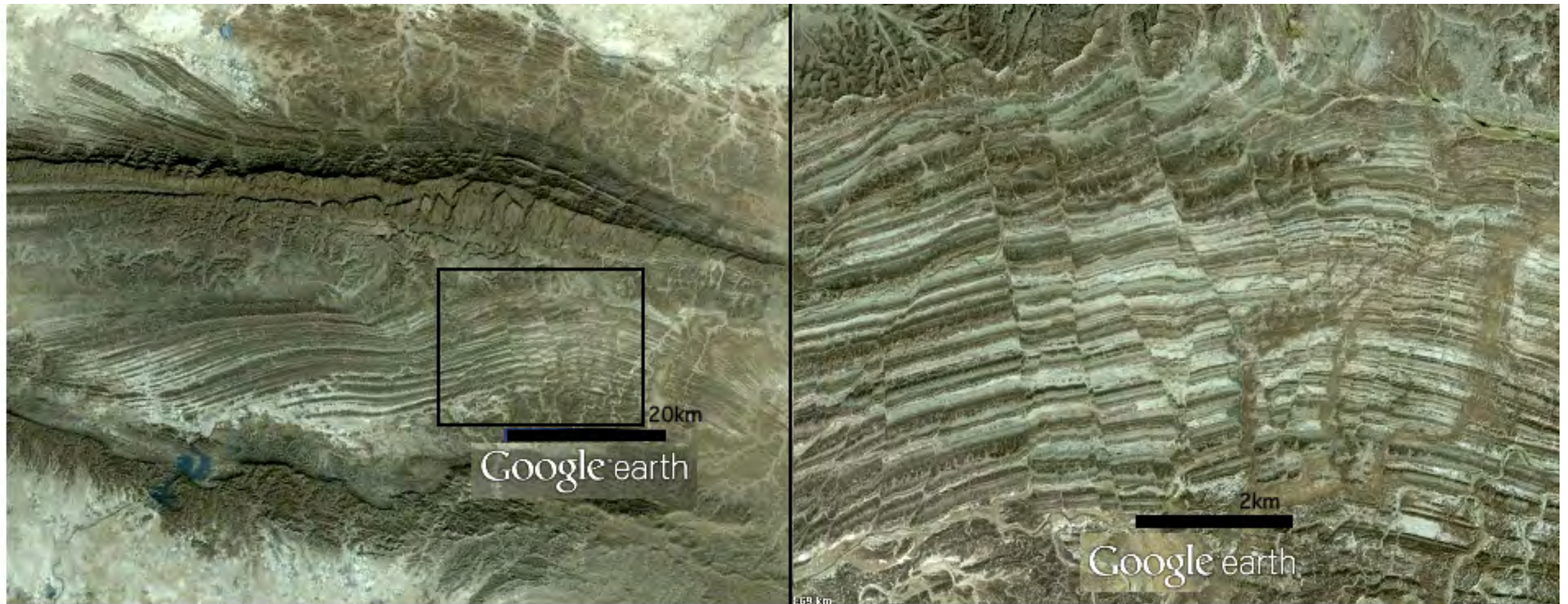◄   **Check Answer**   ►

Fractures and Faults

Fractures and faults are prominent in the upper part of the continental and oceanic lithosphere, where they are often associated with earthquake activity. The above Google Earth images show a fracture field in southern Pakistan (N25º35/E62º10). Open Google Earth and check for yourself how many phases of faulting you can argue for.
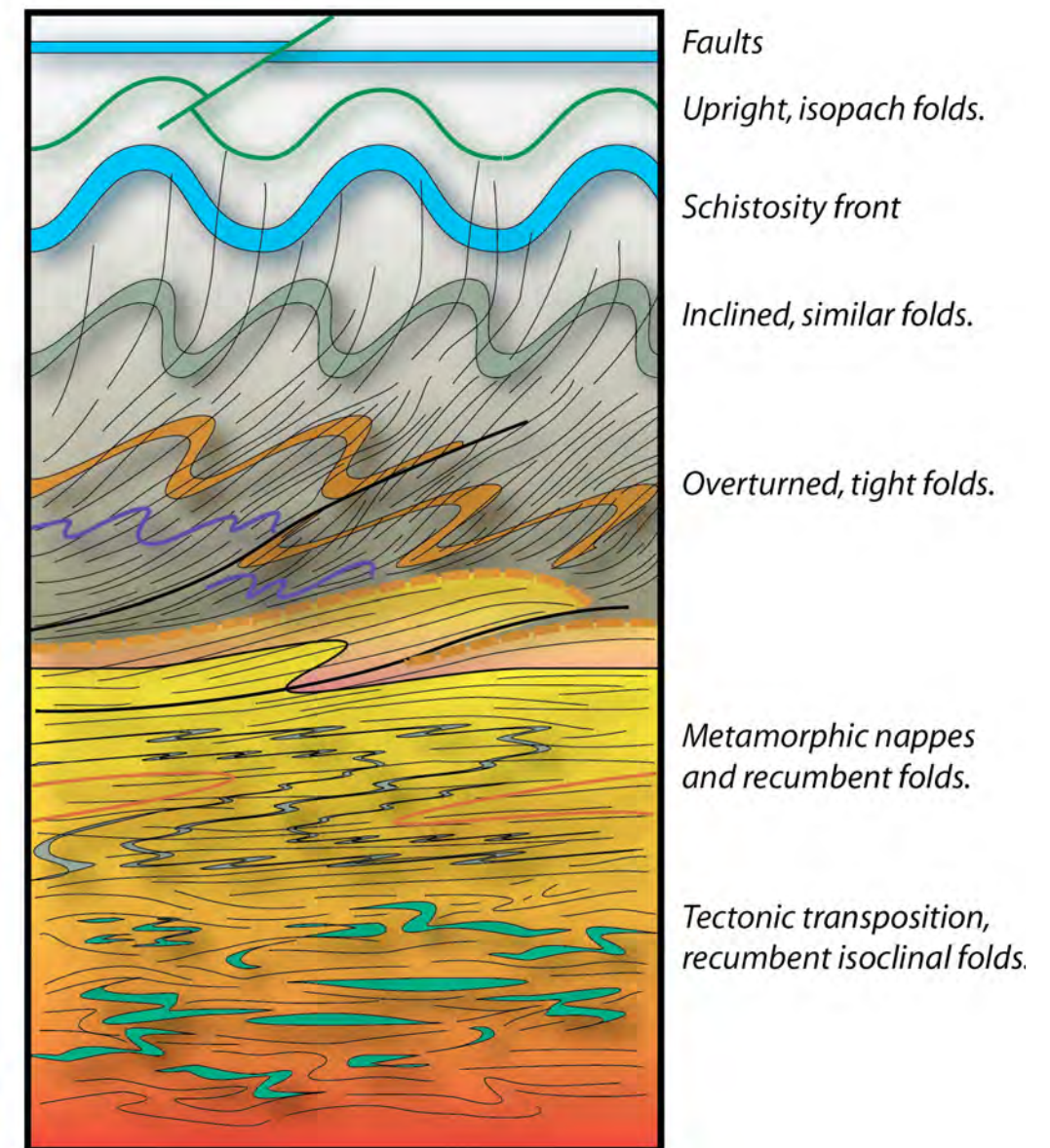
The Concept Of Structural Level

The concept of *Structural Level* is based on the observation that the style of deformation changes with depth due to changes in temperature and pressure (the product between *density*, *depth* and *gravitational acceleration*). With depth, strain becomes more pervasive, synchronous with metamorphism, and often leads to sub-horizontal strain structures and tectonic fabrics such as foliations and lineations.

The **Upper Structural Level** (~0-15 km) is the domain of <u>brittle deformation</u> (faults & fractures), upright isopach folds, extensional fractures and stylolitic joints. Bedding is always recognizable.

The **Intermediate Structural Level** (~10-25 km) starts at the schistosity front, where schistosity (also known as cleavage) begins to form. The style of folding includes: similar folds, tight folds and overturned folds with a strong axial planar schistosity. Pressure solution is the dominant cleavage-forming mechanism. As water is released during prograde metamophism abundant quartz veins form, and <u>Ductile deformation</u> dominates over fracturing. Metamorphic grade does not exceed the greenschist to mid-amphibolite facies (~300 to ~600°C).

In the **Lower Structural Level** (>~20 km) the style of deformation includes: metamorphic nappes, large scale recumbent folds and shallow dipping ductile shear zones. Deformation is intense and pervasive, tectonic transposition makes the mapping of early fabrics difficult if not impossible, fabrics are sub-horizontal. Metamorphic grades ranges from amphibolite facies to granulite facies (>600°C), and partial melting is often present.

Gallery 2.1 Structures and structural level.

Faults

Upright, isopach folds.

Schistosity front

Inclined, similar folds.

Overturned, tight folds.

Metamorphic nappes and recumbent folds.

Tectonic transposition, recumbent isoclinal folds.

Illustrative structural cross-section through the continental crust (from surface to Moho) documenting the concept of structural level. Faults and fractures are prominent in the upper structural level.

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Joints, Stylolites & Stylolitic Joints

Joints: A joint is a planar discontinuity with no relative displacement of the adjacent blocks. Joints develop during the exhumation of rocks following erosion of the overburden. Joints result from contraction and expansion due to cooling and decompression respectively.

Stylolites and **stylolitic joints**: Stylolitic joints are discontinuities that result from a deformation mechanism called "pressure-solution". They form through stress-induced dissolution along an irregular surface. Dissolution is triggered by stress concentration at the contact between grains. This process puts into solution molecules detached from minerals and clastic grains.

Stylolitic joints are often darker than their host rocks as iron-coated insoluble particles accumulate in the joint. In section, stylolitic joints are made of tooth-like cones. The cone axes can be either perpendicular to the joint or at an angle, but they always point in the direction of the maximum stress at the time of their formation.

Columnar jointing in a felsic volcanic. Devil Tower, Wyoming (Ph. Credit: Rosyew)

Extensional Fractures & Shear Fractures

Fractures are discontinuities with limited displacement. They form when applied stress reaches the yielding threshold, i.e. the stress at which rock fractures. There are two types of fractures classified into 3 modes:

•**Extensional Fractures** form when the adjacent blocks move away from each other in a direction sub-perpendicular to the fracture plane. The direction of opening of the fracture is parallel to that of the least resistance (i.e. the least stress). Often, extensional fractures define "en échelon" array or even "conjugated en échelon array". Extensional fractures are also known as "tension gashes". Very often these fractures are cemented with minerals that precipitate from the solution filling the rock's pore space. Quartz in silica-rich rocks and calcite in carbonate-rich rocks commonly seal the fractures.

Extensional fractures and stylolitic joints are often associated and form Mode 1 and anti-Mode 1 fracturing. Together they reveal the orientation of the three principal stress axes. This micro-structure association forms the basis of paleostress analysis, which consists of mapping stress trajectories.
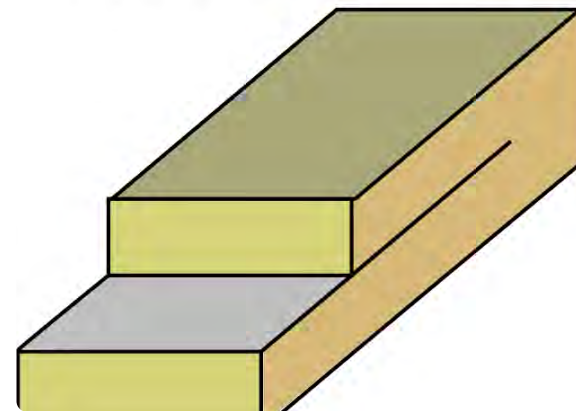
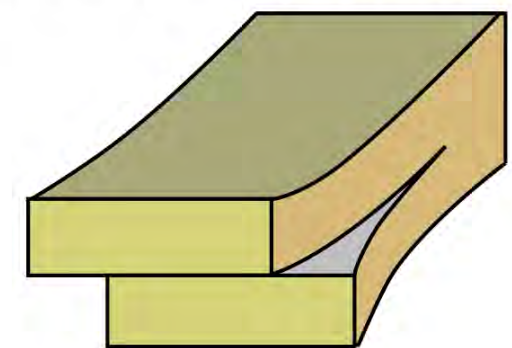*Mode I (Extensional fracture)*   *anti-Mode I (Stylolitic joint)*
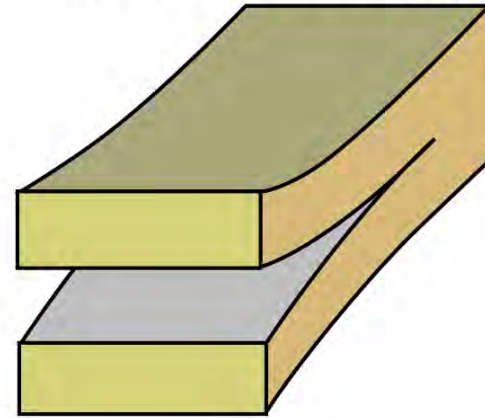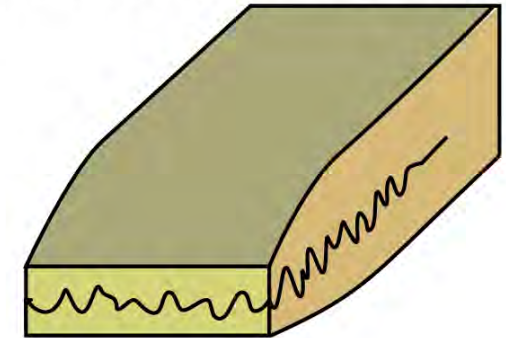
*Mode II (Shear fracture: sliding)*   *Mode III (Shear fracture: tearing)*

Fractures are classified into 3 main modes:

• **Shear Fractures** involve displacement of adjacent blocks parallel to the fracture plane. As adjacent blocks slip along each other, deformation affects the region of contact (i.e. damage zone). Drag folds and small offsets in pre-existing layering in the damage zone help to infer the kinematic of shear fractures. In structural geology we call "kinematic of a fault" or "kinematic of a shear fracture" the relative sense of motion parallel to the discontinuity.

• Shear fractures exist at any scale from microns to hundreds of kilometers, and are often seismogenic (i.e. they are associated with earthquakes).
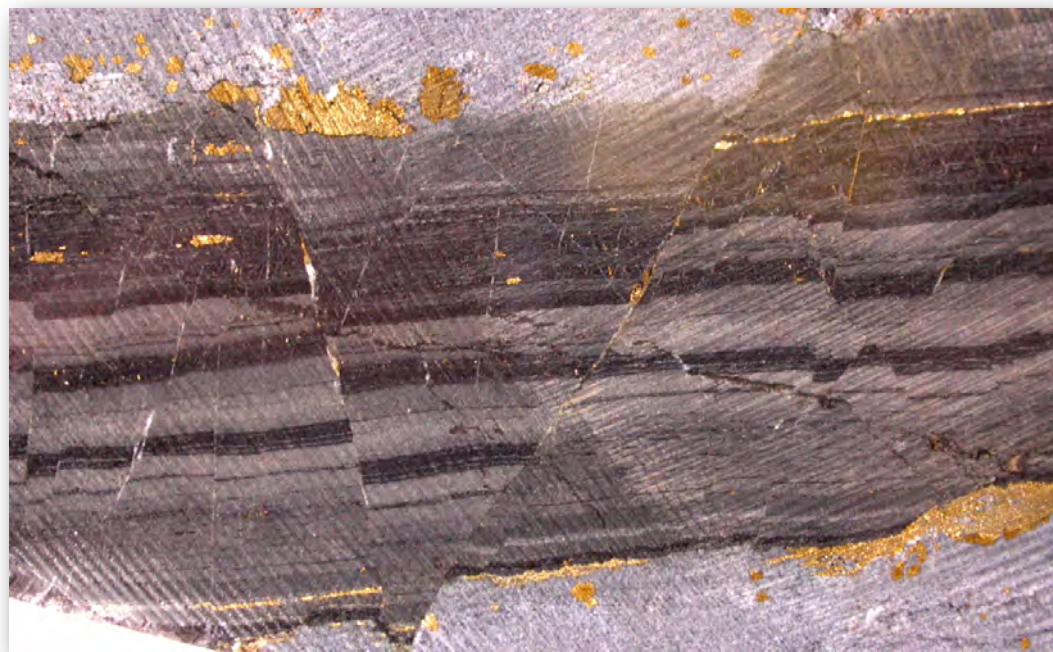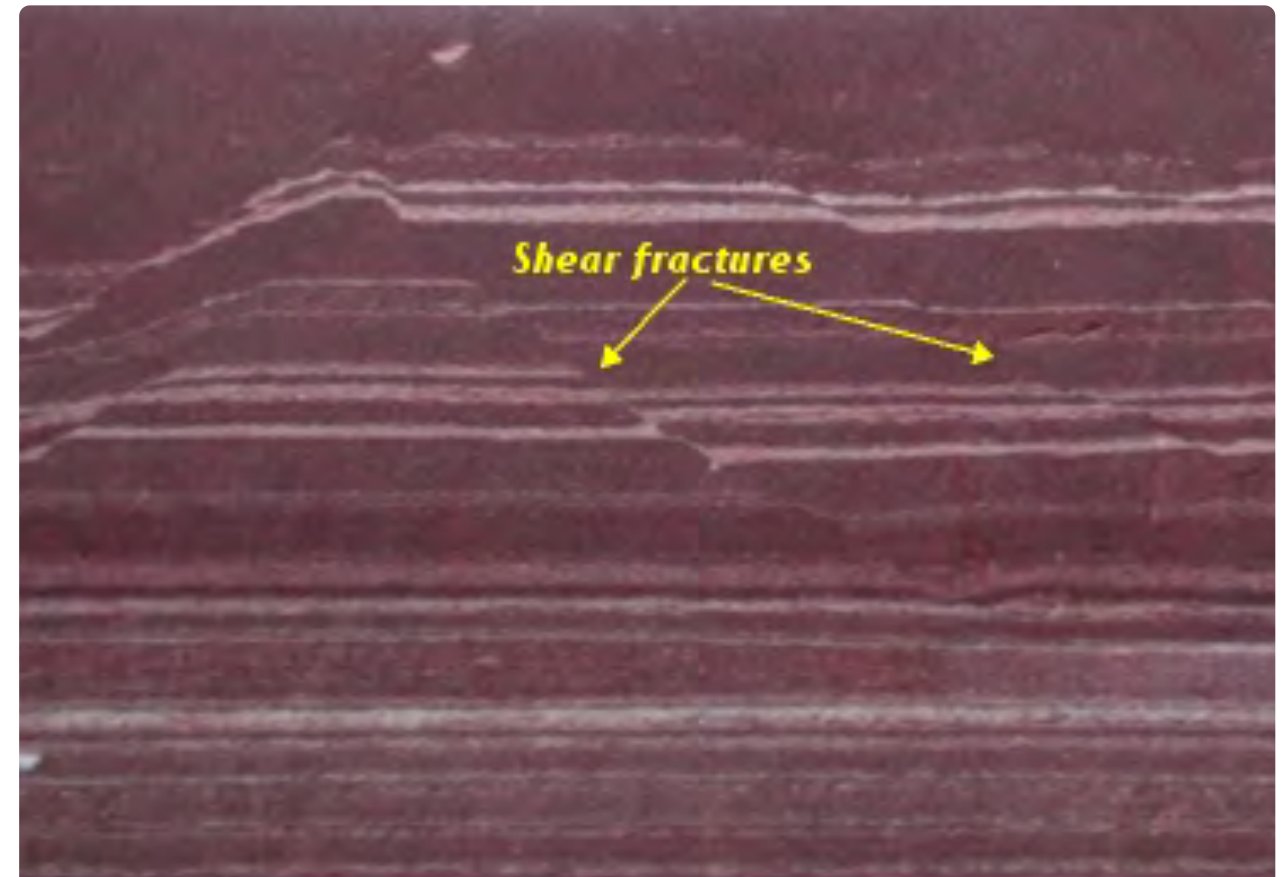
Shear fractures

ed shear fractures. Note the offsets of the layering parallel to planes.

Micro-scale shear fracture in a finely laminated shale (length of the photo is about 4 cm, Photo credit: P. Rey)...

Faults & Tectonic Regimes

•**Faults**: A fault is a fracture across which two blocks have slipped; the displacement of adjacent blocks is parallel to the fault plane. Faulting corresponds to the brittle failure of an undeformed rock formation or, alternatively, involves frictional sliding on a pre-existing fault plane. Faulting occurs when the maximum differential stress (i.e., maximum stress σ_1 minus minimum stress σ_3) exceeds the shear strength of an intact rock formation, or the frictional strength of a pre-existing fault.

•**Fault types**: There are three types of faults.
Normal faults: the hanging wall moves down dip relative to the footwall. Normal faults dominates in the **extensional tectonic regime** (the maximum stress σ_1 is vertical).
Reverse faults: the hanging wall moves up dip relatively to the footwall. Reverse faults are common in the **contractional tectonic regime** (the minimum stress σ_3 is vertical).
Strike-slip faults (dextral or sinistral): the blocks move horizontally past one another. Strike-slip faults often indicate a **transcurrent tectonic regime** (the intermediate stress σ_2 is vertical).
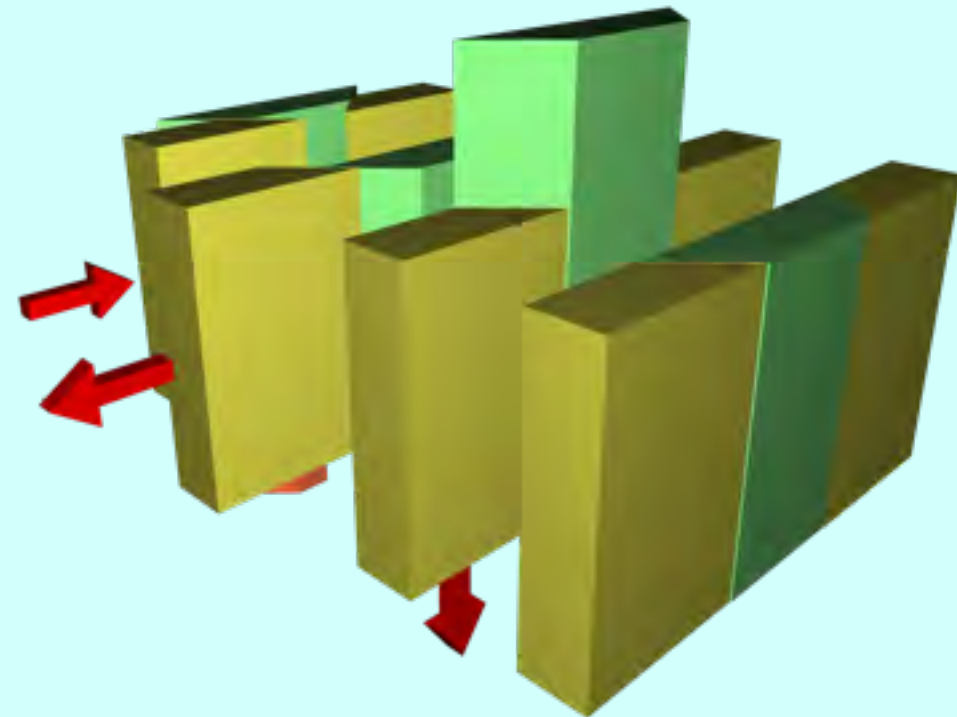
Three tectonic regimes.

These sketches illustrate various examples of faults. There are the 3 simple end-member cases of normal, reverse and strike-slip faults. The geometry of faults can however be a bit more complicated. The slip direction can be oblique to the strike of the fault (e.g., fault kinematic can be either normal sinistral, normal dextral, reverse sinistral or reverse dextral). The offset magnitude can vary along strike, the fault is said to be rotational and described as a *scissor fault*. Finally fault planes are not always planar surfaces and can be curved. A listric fault has a dip that decreases with depth.

At the Earth's surface, one of the principal stress axes is always vertical because air is too weak to sustain a shear stress component. Hence, at the Earth's surface the stress regime is said to be **Andersonian**. It may come as a surprise that a fault with a slip direction oblique to the strike (for instance a normal sinistral fault) can form close to the surface. In fact, it can be shown (Bott, 1959) that for a given Andersonian state of stress (i.e. state of stress with one vertical principal stress axis) the orientation of the slip direction (i.e the striea) can have any orientation on a *pre-existing* fault plane, and this orientation is only a function of the relative magnitude of the 3 principal stress axes.
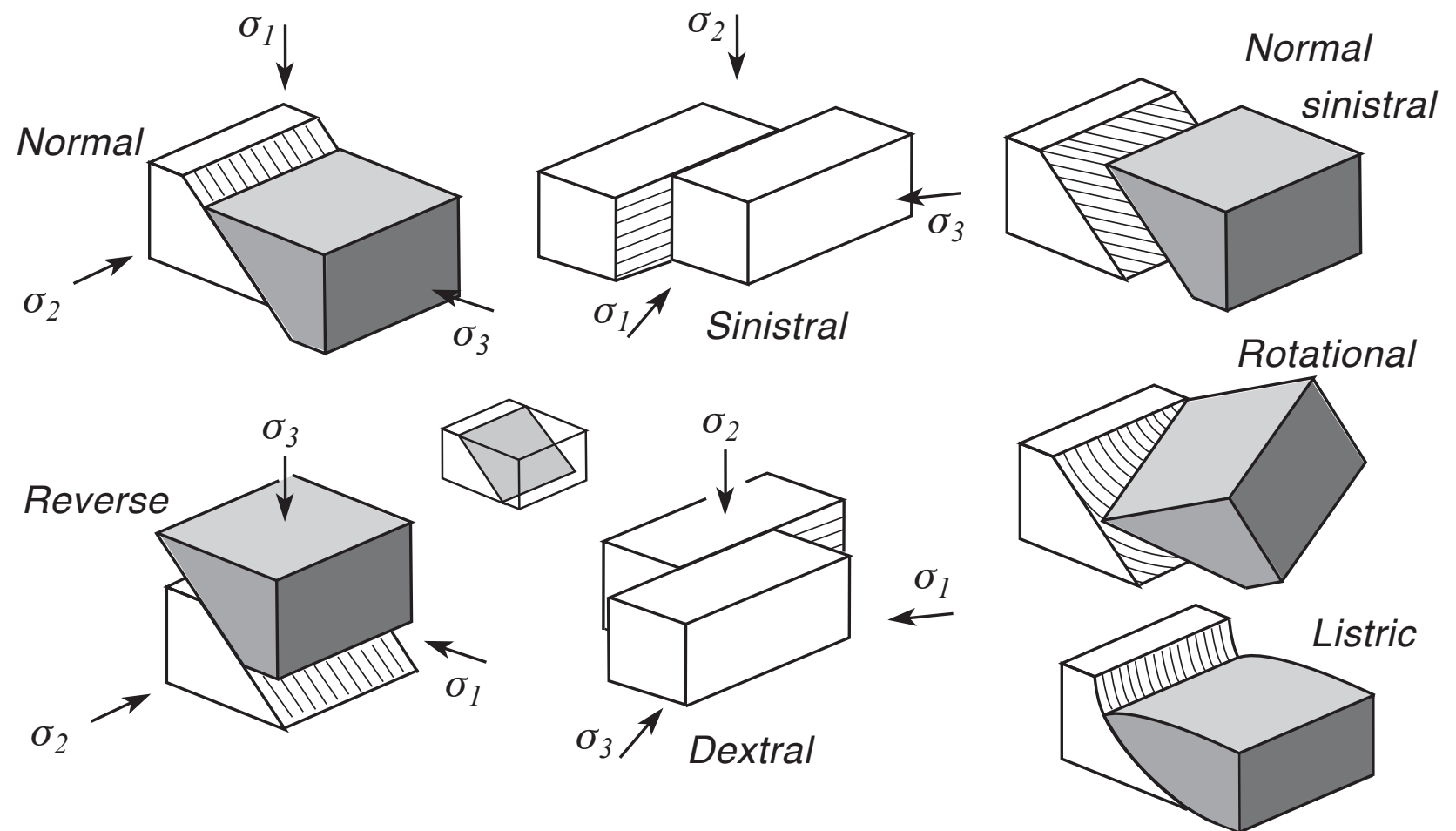
Measuring Fault Slip

Structural geologists are interested in "fault kinematic". In other words they want to find out the relative sense of motion of blocks on either side of faults, as well as the orientation and amplitude of the slip. Fault slip is characterized by a vector called the net slip. Its direction is that of the slip and its length is the amplitude of the slip. On a fault plane the direction of the net slip is often given by striae (i.e. a friction mark). The net slip can be decomposed into either 1/ two orthogonal components on the fault plane (strike-slip and dip slip components); or 2/ two orthogonal components in the vertical plane that contains the net slip vector (horizontal heave and vertical throw), or 3/ three orthogonal components in the geographic system (strike slip, heave, and vertical throw).

The "kinematic plane" is the plane which is perpendicular to the fault plane and parallel to the net slip vector (ie the striae). For *newly* formed faults the kinematic plane contains the principal stress axis σ_1 and σ_3 with σ_1 at 30° from the fault plane.
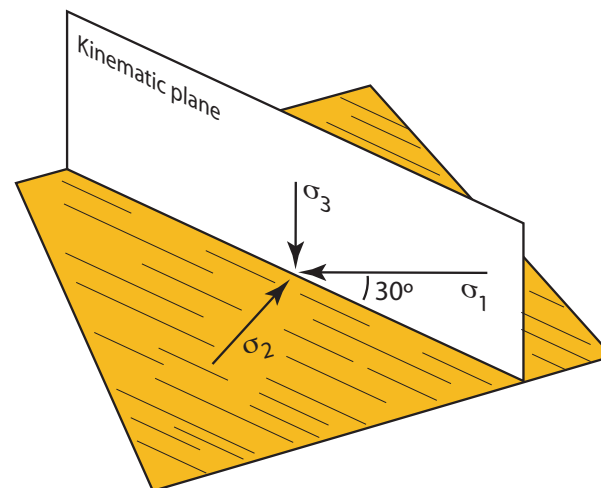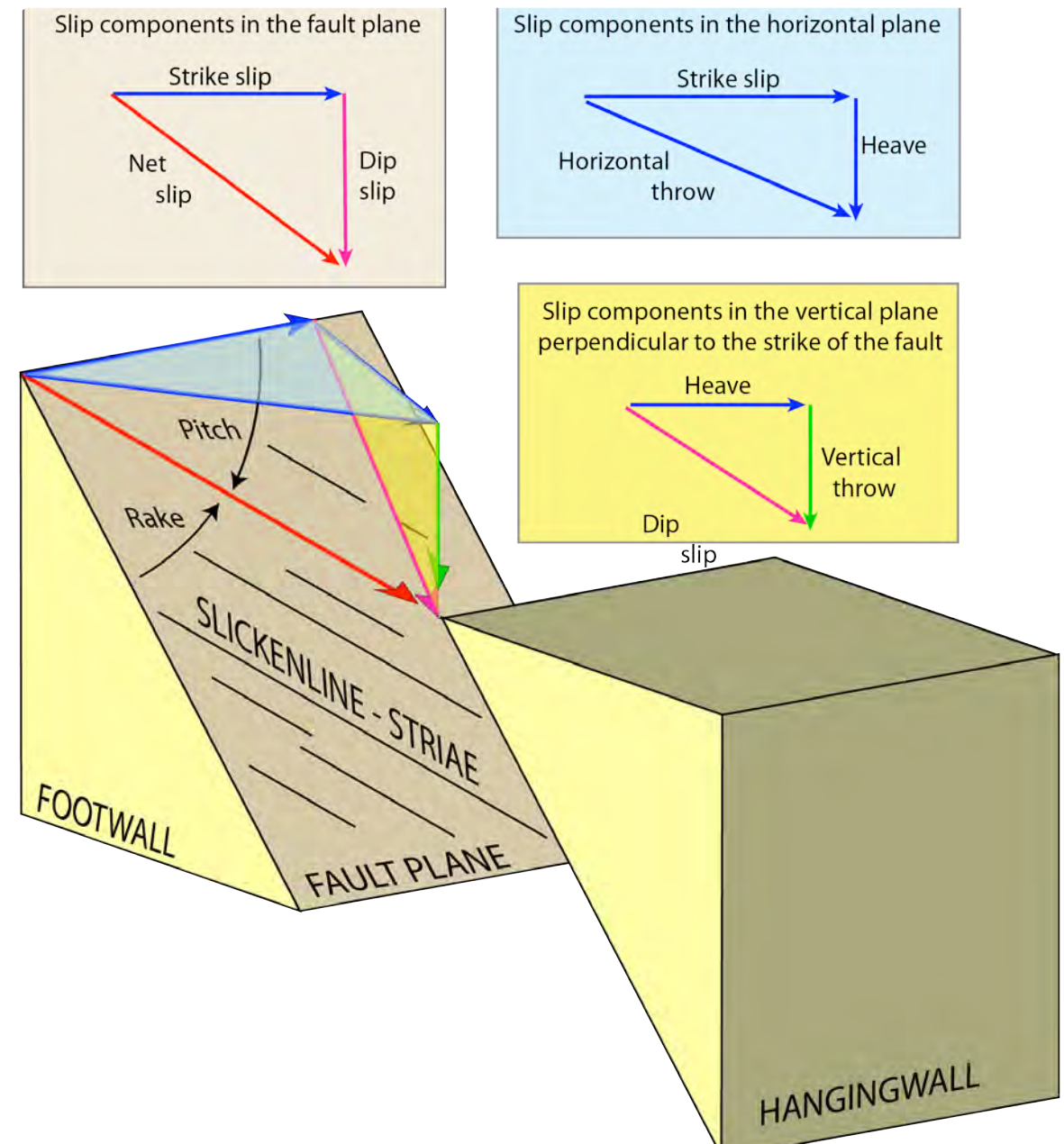
*The net slip vector and its components.*

Fault Kinematic Analysis

Kinematic analysis is the art of figuring out the relative sense of motion across faults and shear zones.

To do this, structural geologists use "kinematic criteria" i.e. structures and microstructures that develop during faulting and whose geometric characteristics depend on the sense of shear.

The main kinematic criteria for brittle faults are:

1/ **Riedel shear fractures** (R and R'): R are synthetic at 10-15° to the fault plane; R' are antithetic (opposite sense of shear) and oriented at 70-80° to the fault plane.

2/ **Tool marks**: Cavities formed by the mechanical erosion of the fault plane by hard objects.

3/ **Extensional fractures** at ca. 30° angle to the fault plane (also called **T fractures**).

4/ **Mineralized steps**: Step-shaped cavities filled with fibrous minerals (often quartz or calcite).

5/ **Dry steps**: same as above but with no crystallization.

6/ **Rough surfaces** (usually with *stryloliths*) form due to pressure solution, and **smooth surfaces** (polished) .
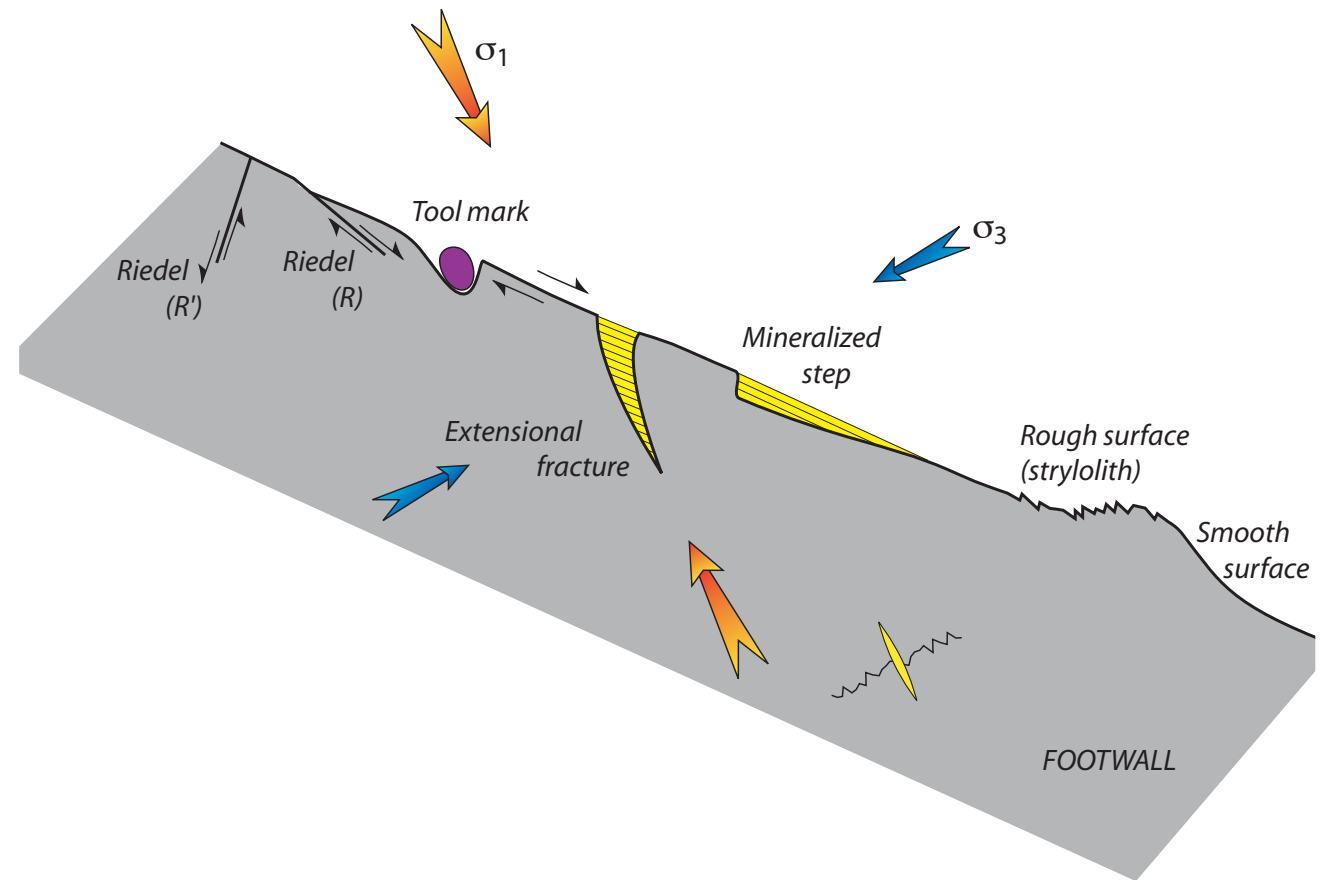
Kinematic criteria shown on a cross-section perpendicular to a fault plane, the plane of the cross-section contains the striae (i.e. the plane of the cross-section is parallel to the kinematic plane). Only the footwall block is shown.

A Little Quiz …

1/ Define the following terms: a fact, an hypothesis, a model, a theory and a law.

2/ Define the Darwinian scientific method.

3/ A scientific model should always verify a number of characteristics. What are they ?

4/ How do stylolitic joints form?

5/ What are the two main types of fracture?

6/ Draw a stylolite / extensional fracture association and give the orientation of the principal stress axes.

7/ What are the three main types of faults?

8/ Describe the three modes of fractures?

9/ Define the following concepts: a/ A listric fault. b/ The kinematic plane.

10/ Define the following concepts: Vertical and horizontal throw, the heave, the net slip?

The picture below shows a cemented breccia, it is the result of faulting. The rock crumbled into numerous angular clasts. A silica-rich fluid circulated through the fault zone precipitating quartz that cemented the clasts together.

A LITTLE BRAIN TWISTER ...

N

50

25

Produced with http://www.visible-geology.appspot.com/
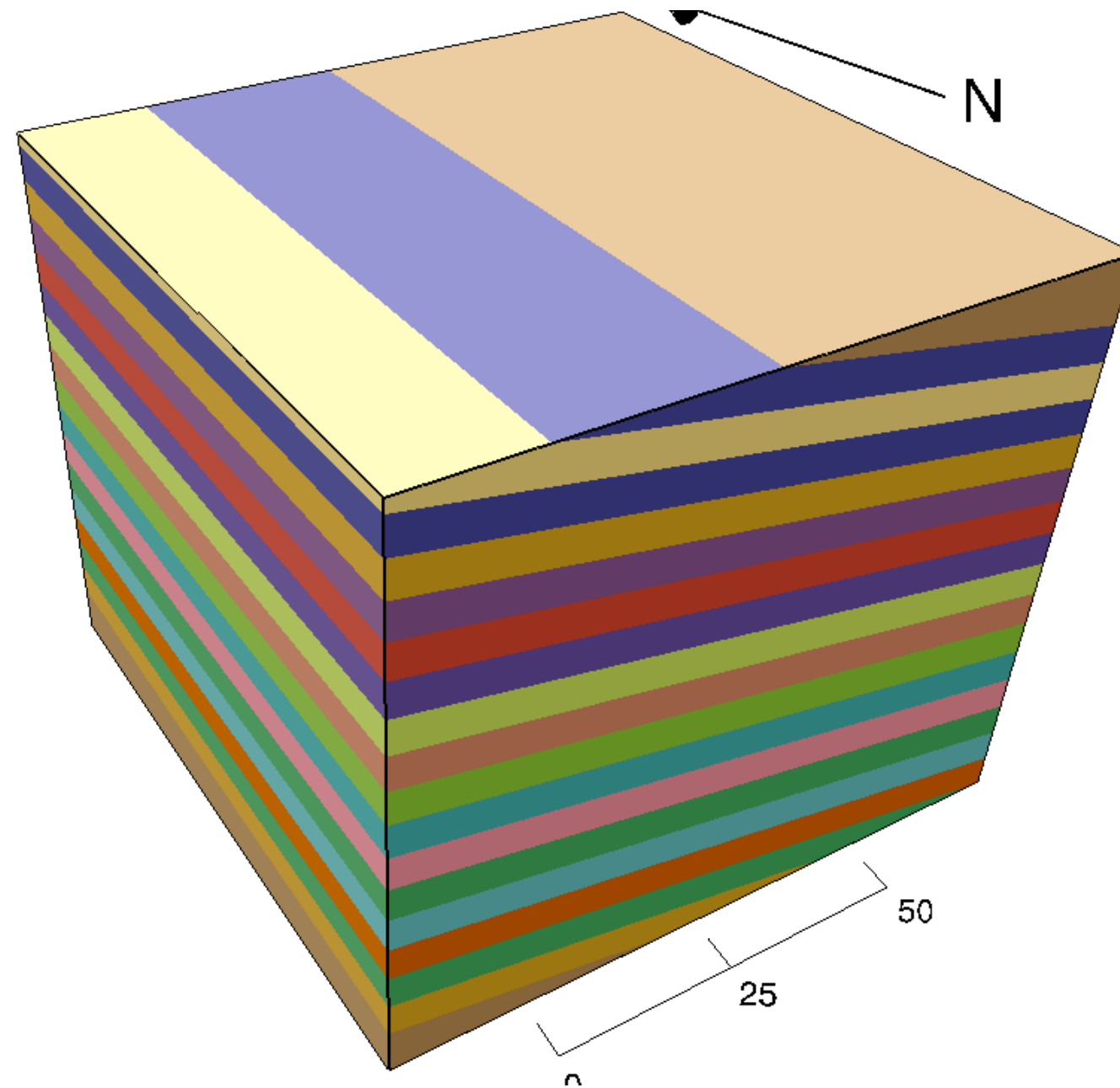
A sedimentary rock formation is gently tilted to the east...

24

Question 1 of 7

The concept of *"structural level"* refers to:

○ **A.** The level of structural complexity.

✓ **B.** The partitioning of strain style with depth.

○ **C.** The boundary between brittle deformation and ductile deformation.
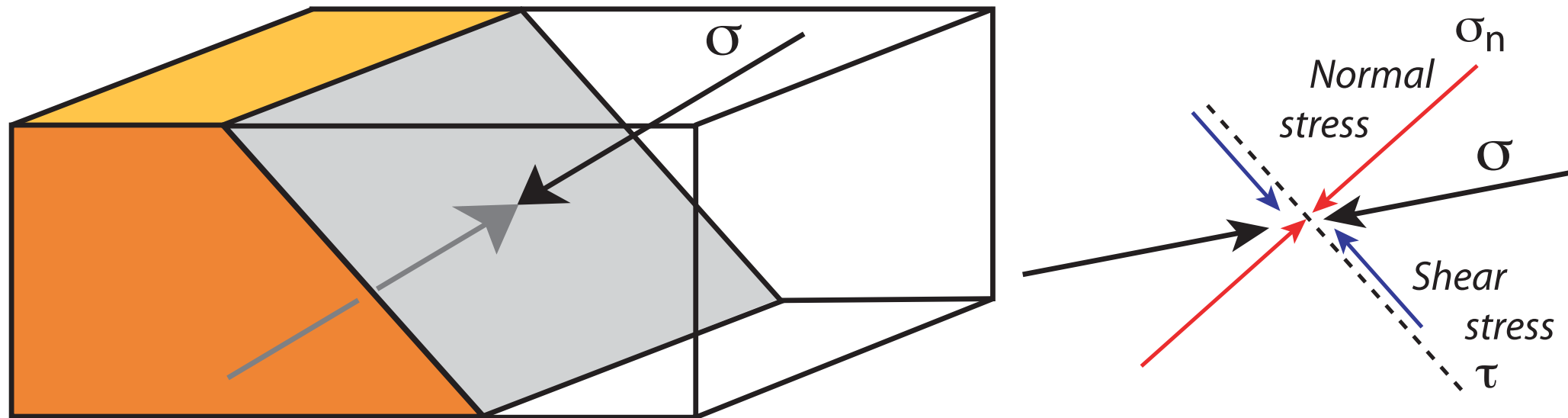
○ **D.** The number of phases of deformation.

◀   **Check Answer**   ▶

Force, Traction and Stress

Stress: a couple of traction acting across a surface

We have seen in the former chapter that the orientation of newly formed and reactivated faults and fractures can be easily used to infer the orientation of the principal stress axes responsible for faulting. At this point we need to clarify the concepts of force, traction and stress. Forces, tractions, and stresses are the causes of deformation.

The Concepts Of Force & Traction

According to Newton's second law of motion, a force is a physical entity which tends to change either the state of rest or the uniform motion in a straight line of a body. A force is described by a vector, its SI unit is the Newton (N). Its magnitude is equal to the rate of change of momentum of the body. The force F producing the acceleration a (m . s^{-2}) of a body of mass m (kg) is therefore given by: $F = m.a$. Hence, a force of 1 Newton accelerates a 1 kg object by 1 m per second per second.

There are two types of forces:

Surface forces act on the surface of a body. For instance the tectonic forces (e.g. ridge push and slab pull) acting on a lithospheric plate are surface forces, they are often called external forces.

Volume forces (or body forces) act inside the body, they are often called internal forces. In a lithospheric plate, body forces arise due to gravity pulling every molecule toward the center of the Earth (P), and heterogeneities in density (e.g. lateral density changes due to crustal thickening).

The concept of **Traction**: In mechanics, traction is a pressure vector that represents the limit of the ratio of force over area as area tends to zero. This force can have any orientation with respect to the surface. Traction is represented by a vector whose orientation is that of the force and its magnitude the ratio of the force (N) over the area (m^2) of the surface. The unit of a traction is therefore that of pressure: N.m^{-2}. Tractions can be decomposed into a shear component (parallel to the surface) and normal component (normal to the surface).
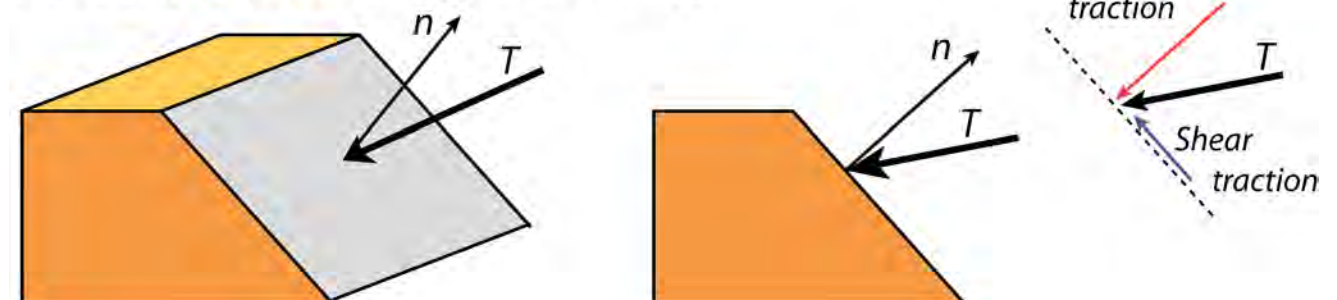

Surface force

Volume force
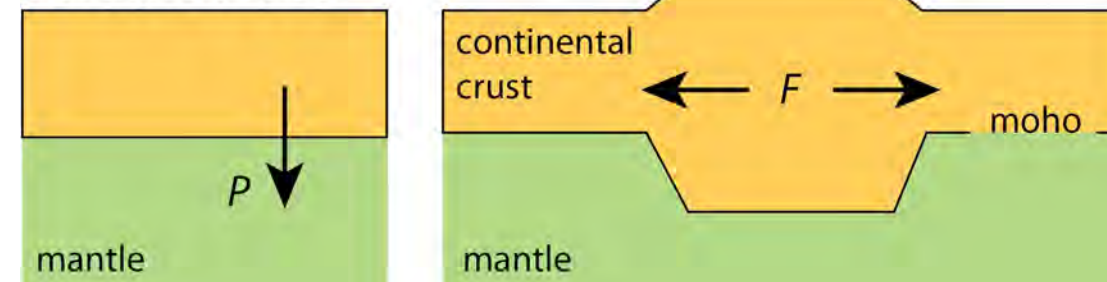
*Traction: A force acting on a surface*

The Concept Of Stress

In a volume of rock, let's consider a physical point at rest, and let's consider a small surface centered on that point. The orientation of the small surface is characterized by its "normal" n (i.e. the vector orthogonal to the surface, the length of which is proportional to the surface area).

The stress acting across this surface is defined by the couple of traction vector acting on opposite sides of the surface. In a strict sense, unbalanced tractions acting on a surface do not give rise to stress; they cause acceleration. Hence, the concept of stress applies to a body at equilibrium, - in a state of rest or uniform rectilinear motion.

Because the surface is at rest, these tractions have same magnitudes (i.e. length) but opposite directions. However, the magnitude and orientation of the stress change with the orientation of the small surface.

If one could record the orientation and length of the stress as the surface rotates about its centre, one would see that the end tips of the stress define an ellipsoid. The **stress ellipsoid** is characterized by three principal orthogonal stresses (i.e. 3 pairs of equal but opposite tractions) σ_1, σ_2 and σ_3 with $\sigma_1 \geq \sigma_2 \geq \sigma_3$. There are only three positions of the surface for which the stress is perpendicular to it, and that σ_1, σ_2 and σ_3 are the **normal stress vectors**.
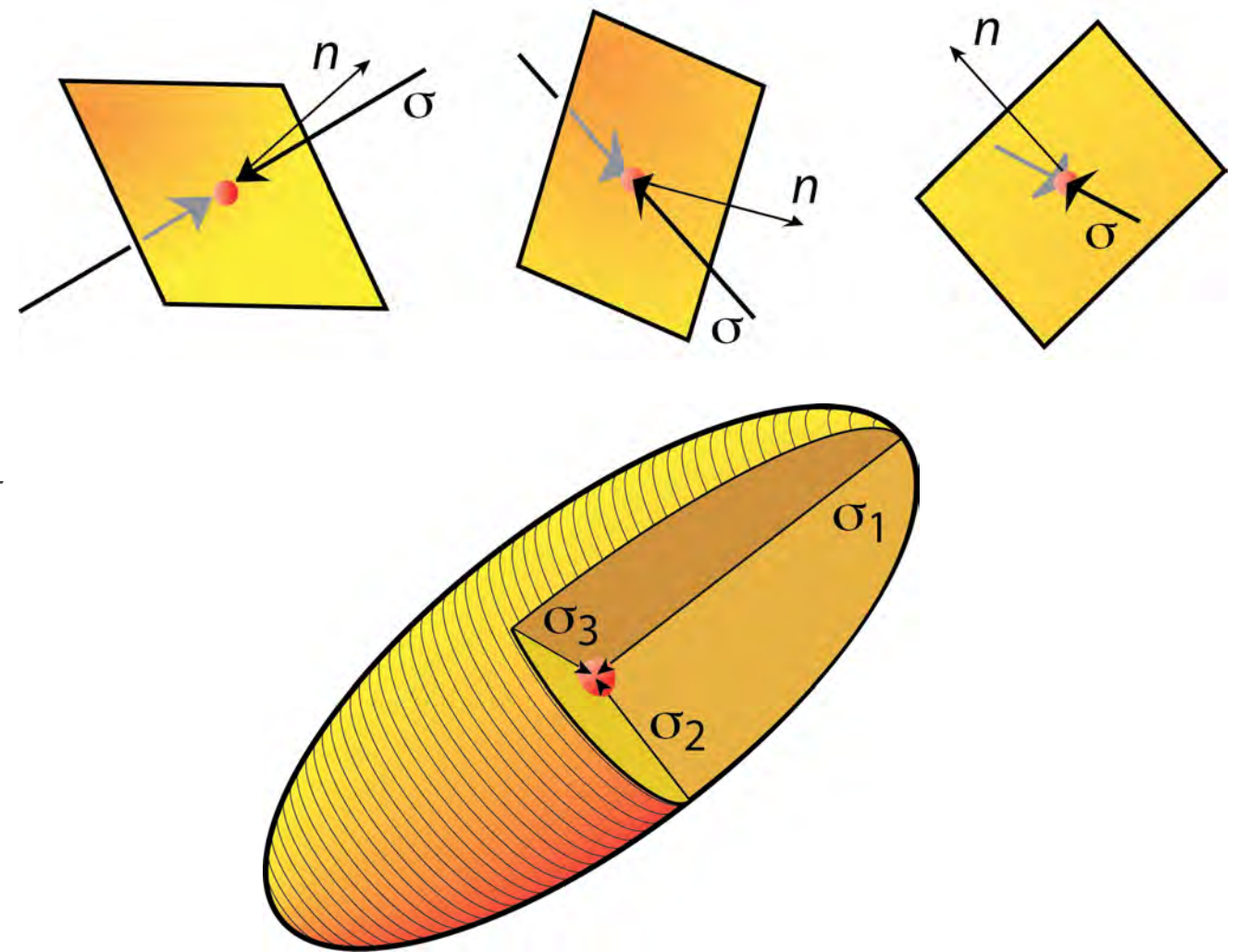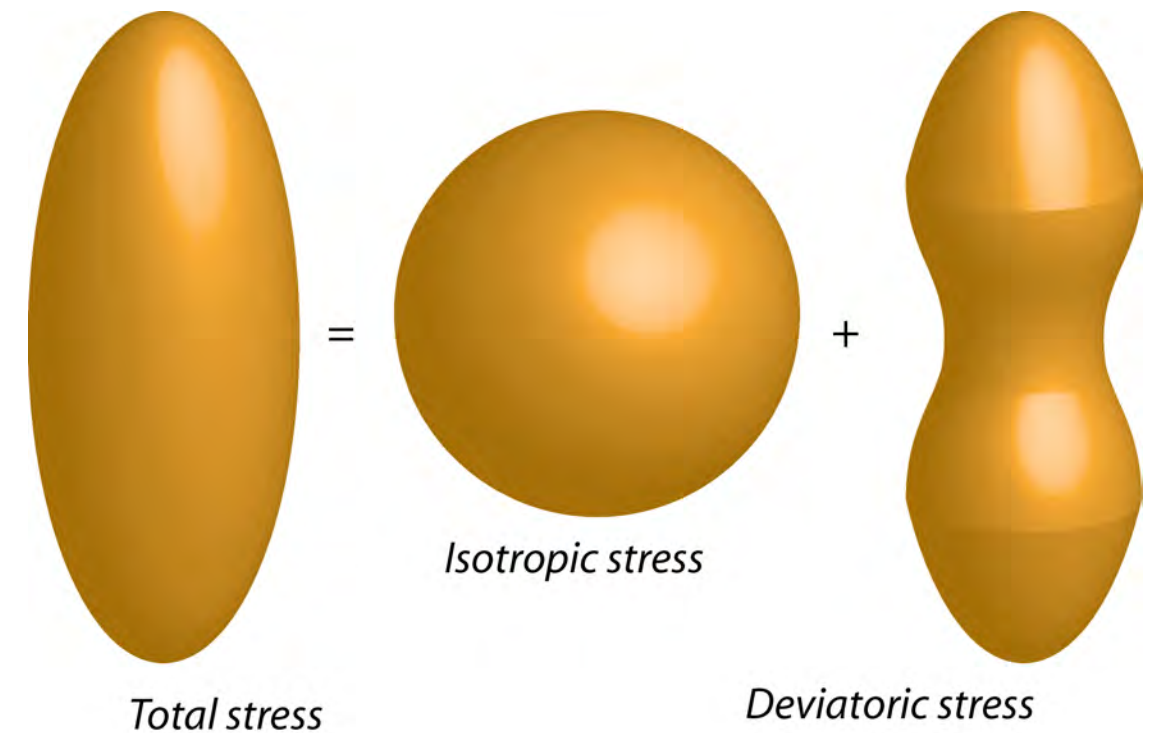
The state of stress acting on a material point is characterized by an ellipsoid called the stress ellipsoid. It is defined by three perpendicular axes called the principal stresses: σ_1, σ_2 and σ_3. All stresses are positive (i.e. directed toward the material point). Extensional structures result from material moving or flowing in the direction of least stress.

At the Earth's surface the state of stress is *Andersonian*, which means that one of the principal stress axes is vertical.

The state of stress acting on any point in the Earth is characterized by a stress ellipsoid. The simplest stress ellipsoid is a sphere: $\sigma_1=\sigma_2=\sigma_3$. In such a case, the stresses are the same in all directions and the state of stress is **isotropic**. Such a state of stress is also said to be **hydrostatic** or **lithostatic**. The magnitude of the stress (i.e., the radius of the sphere) is simply the pressure due to the weight of the column of rock above the considered point. This pressure, often called the confining pressure, comes with the same intensity in all direction and its magnitude is given by: $\rho.g.z$, where ρ is the density, g is the gravitational acceleration and z is depth. Lithostatic pressure does not drive deformation besides small amount of compaction or expansion.

When the state of stress is **anisotropic**, the stress ellipsoid can be decomposed into an **isotropic stress** component of magnitude $\sigma_i = (\sigma_1+\sigma_2+\sigma_3)/3$, and a **deviatoric stress** component that accounts for the difference between the **total stress** and the isotropic stress. The deviatoric stress component is responsible for deformation.

Mathematically, the state of stress is a 2nd order tensor (9 components). This tensor is symmetric ($\sigma_{ij}=\sigma_{ji}$) and the sum of its diagonal components ($\Sigma\sigma_{ii}$=constant) is invariant (the value of each component changes as one changes the coordinate system, however $\Sigma\sigma_{ii}$ is independent of the coordinate system). This tensor can be decomposed into an isotropic component and a deviator. It is always possible to choose a coordinate system so that the tensor is reduced to its principal stress components: i.e.: $\sigma_{ij}=0$, and $\sigma_{11}=\sigma_1$, $\sigma_{22}=\sigma_2$, $\sigma_{33}=\sigma_3$.

Total stress = Isotropic stress + Deviatoric stress

$$\sigma_{ij} = \begin{bmatrix} \sigma_{11} & \sigma_{12} & \sigma_{13} \\ \sigma_{21} & \sigma_{22} & \sigma_{23} \\ \sigma_{31} & \sigma_{32} & \sigma_{33} \end{bmatrix} = \begin{bmatrix} \sigma_i & 0 & 0 \\ 0 & \sigma_i & 0 \\ 0 & 0 & \sigma_i \end{bmatrix} + \begin{bmatrix} \sigma_{11}-\sigma_i & \sigma_{12}-\sigma_i & \sigma_{13}-\sigma_i \\ \sigma_{21}-\sigma_i & \sigma_{22}-\sigma_i & \sigma_{23}-\sigma_i \\ \sigma_{31}-\sigma_i & \sigma_{32}-\sigma_i & \sigma_{33}-\sigma_i \end{bmatrix}$$

Stress tensor = Isotropic stress component + Deviator

$\sigma_{ij} = \sigma_{ij}$
$\Sigma\sigma_{ii}$ = Invariant

$\sigma_i = \dfrac{\sigma_1 + \sigma_2 + \sigma_3}{3}$
Isotropic stress

$\sigma_{ii}-\sigma_i$
Principal differential stresses
Σ Principal Differential stresses = 0

The stress acting on a pre-existing fault plane is represented by a couple of traction vector having similar magnitudes (i.e. length) but opposite directions. For convenience, this stress can be decomposed into a **shear stress component (τ)**, and **normal stress component (σ_n)**. Both are represented by a couple of vector. The normal stress tends to prevent faulting by pushing both blocks towards each other. In contrast, the shear stress promotes faulting.

From experiments we observe that for a newly formed fault (i.e. developed from a fracture-less volume of rock):

1/ σ_1, n and σ_3 belong to the same plane. This plane is called the **kinematic plane**.

2/ Shear stress and normal stress also belongs to the kinematic plane.

3/ The **slip vector** on the fault plane is parallel to the shear stress.

4/ The angle between the fault plane and σ_1 is close to 30°.

The stress is decomposed into normal stress and shear stress...

When pre-existing faults exist, these faults may be reactivated when stressed. In the case of fault reactivation, one can observe from experiments that for a given fault orientation, and for a given state of stress defined by σ_1, σ_2 and σ_3 :

1/ The slip vector on the fault plane (i.e. direction parallel to the striae) is parallel to the shear stress.

2/ The orientation of the shear stress depends on the orientation of n, σ_1, σ_2 and σ_3 and $R = (\sigma_1 - \sigma_2) / (\sigma_2 - \sigma_3)$

3/ It can be shown that for a state of stress stable in orientation, the striae can take any orientation on the fault plane depending on R **(Bott, 1959)**.

Therefore, one can infer from the processing of a large number of faults/striae/kinematic data, the orientation in space of the stress ellipsoid responsible for faulting, as well as the ratio R of the stress ellipsoid.

Introduction To Paleostress Analysis

The determination in the field of the orientation in space of the **paleostress ellipsoid** responsible for the development of stylolites, fractures and faults is relatively easy. We have seen previously that there is a simple set of relationships between the orientation of the paleostress ellipsoid and the orientation of stylolitic joints, shear fractures and extensional fractures that develop in response to an applied stress.

The sketch on the top right shows a set of microstructures that develop due to a tectonic regime in compression. From a stress point of view, a **compressional tectonic regime** is characterized by σ_3 being the vertical stress axis. From a strain point of view it is characterized by vertical stylolitic joints and horizontal extensional fractures.The stylolitic cones point in the direction of σ_1. The normal of extensional fractures is parallel to σ_3. The intermediate stress σ_2 is perpendicular to both σ_1 and σ_3.

In an **extensional tectonic regime** σ_1 is vertical. Stylolitic joints are therefore horizontal, whereas extensional fractures are vertical. Note that horizontal stylolitic joints and vertical extensional fractures can also develop during the compaction of sediments during burial.

In a **transcurrent tectonic regime** σ_2 is vertical. Therefore, both stylolitic joints and extensional fractures are vertical. The stylolitic cones point in the direction of σ_1, the normal to the extensional fracture being parallel to σ_3.

COMPRESSION
- σ_3 Vertical
- Reverse faults
- Vertical stylolitic joints
- Horizontal extensional fractures

EXTENSION
- σ_1 Vertical
- Normal faults
- Horizontal stylolitic joints
- Vertical extensional fractures

TRANSCURRENT
- σ_2 Vertical
- Strike-slip faults
- Vertical stylolitic joints
- Vertical extensional fractures
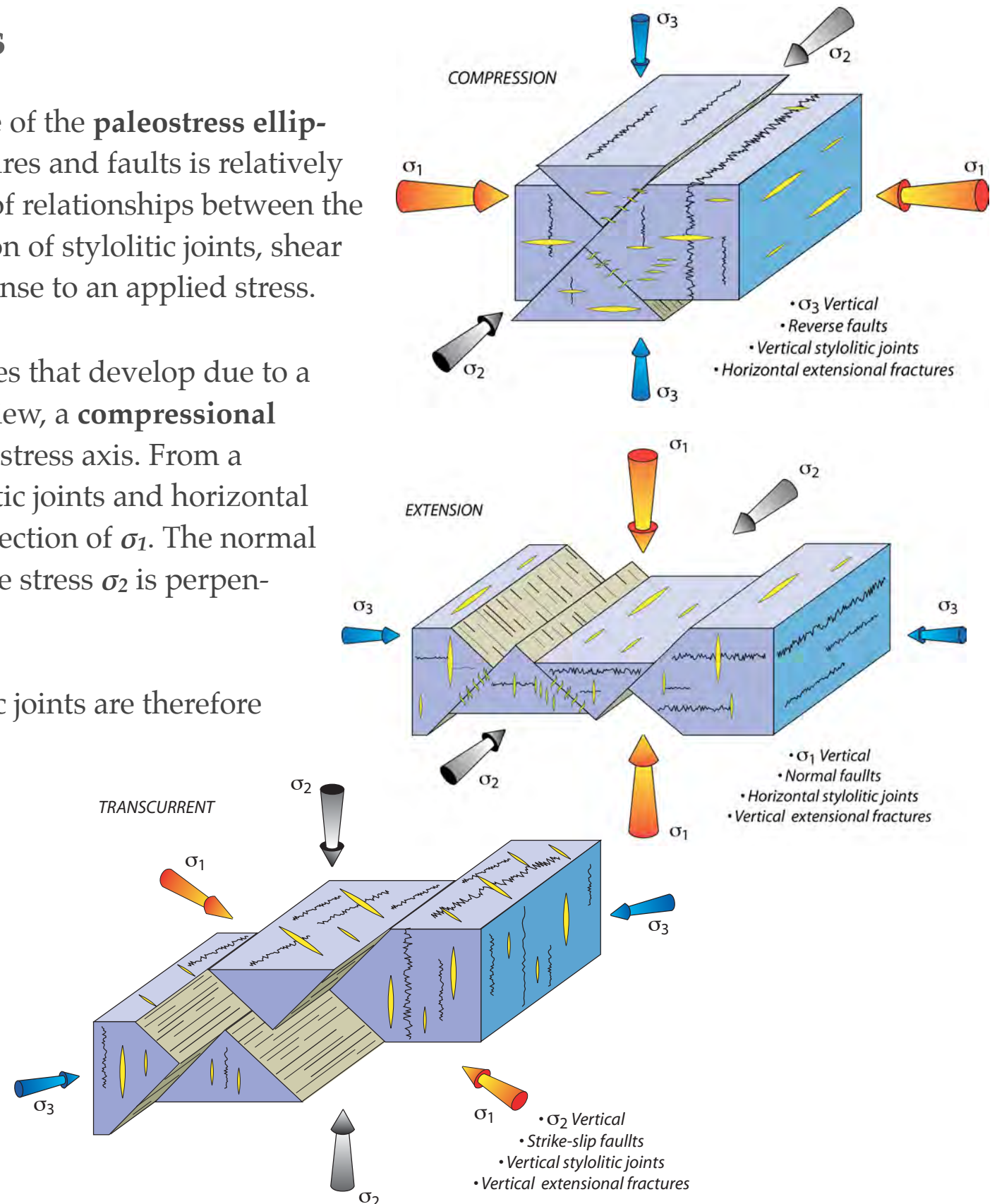
Stress Perturbation

Stress is transmitted through rocks without perturbation as long as the material through which stress is transmitted is mechanically isotropic. This is rarely the case in nature. Hence, the stress field changes, in both magnitude and orientation, in the vicinity of mechanical heterogeneities such as faults.

The sketch on the right illustrates that point. The progressive rotation of both extensional fractures and stylolitic joints show that σ_1 (red trajectories) and σ_3 (blue trajectories) rotate around σ_2 in the vicinity of the fault.

Fault overlaps can be responsible for drastic stress perturbations in the region where faults overlap. The sketch below illustrates this point.

map view

Map view
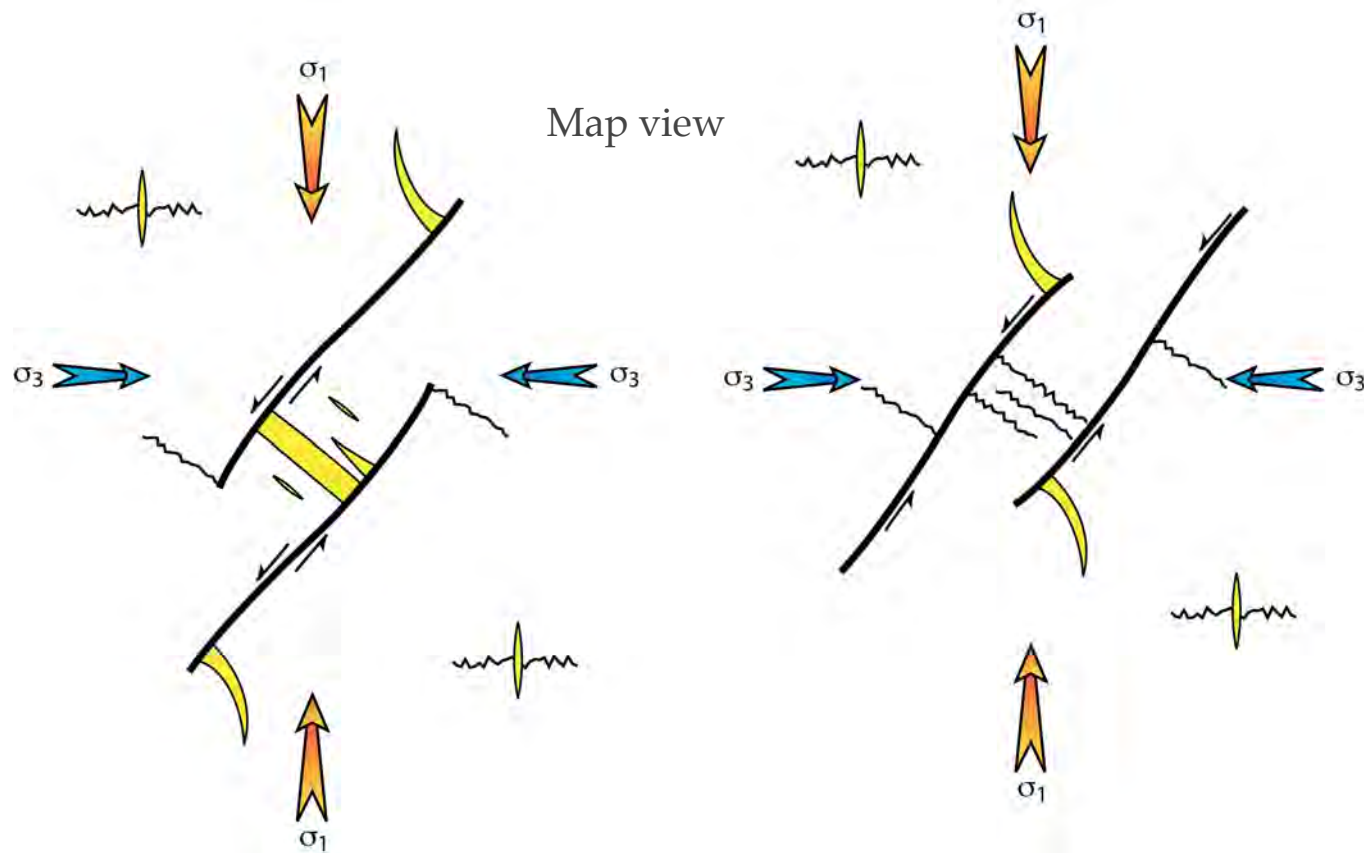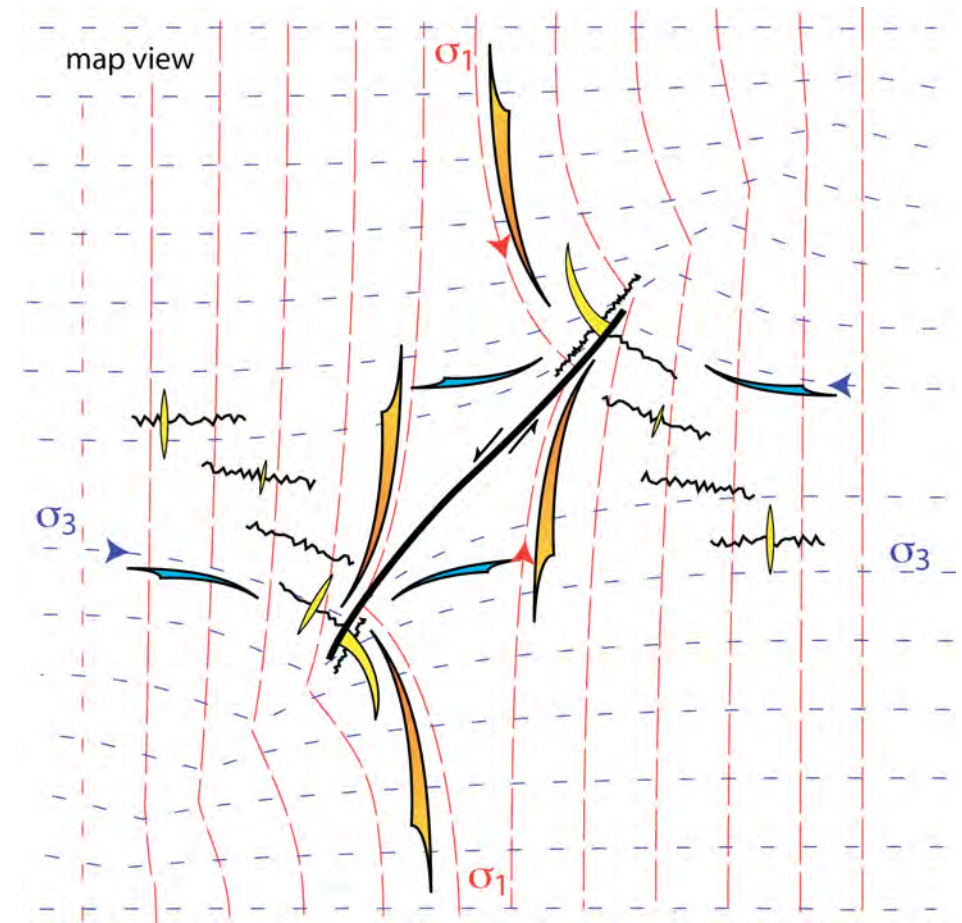
A Little Quiz ...

1/ Define the concepts of force, traction and stress.

2/ The stress acting on a surface can be decomposed into two components. What are they?

3/ The stress on a material point is represented by an ellipsoid. Describe this ellipsoid when the only source of stress is the weight of rocks?

4/ What is the deviatoric stress?

5/ How is σ_i mathematically defined?

6/ How are stylolitic joints oriented in a compressional tectonic regime?

7/ How are extensional fractures oriented in an extensional tectonic regime?

8/ How is the stress ellipsoid oriented in a transcurrent tectonic regime.

9/ What causes stress field perturbation?

Interesting place to visit: http://www.world-stress-map.org
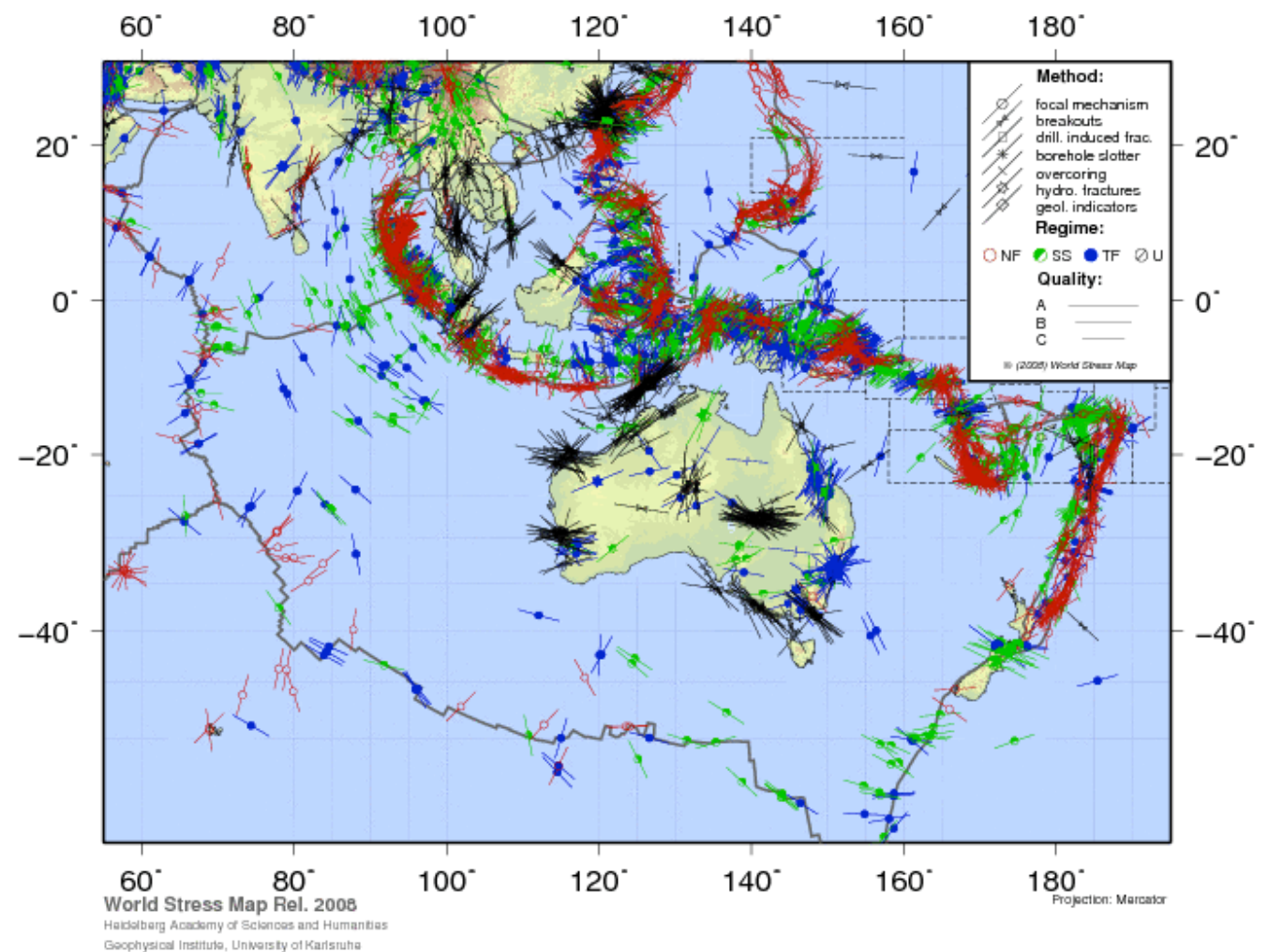
World Stress Map Rel. 2008
Heidelberg Academy of Sciences and Humanities
Geophysical Institute, University of Karlsruhe
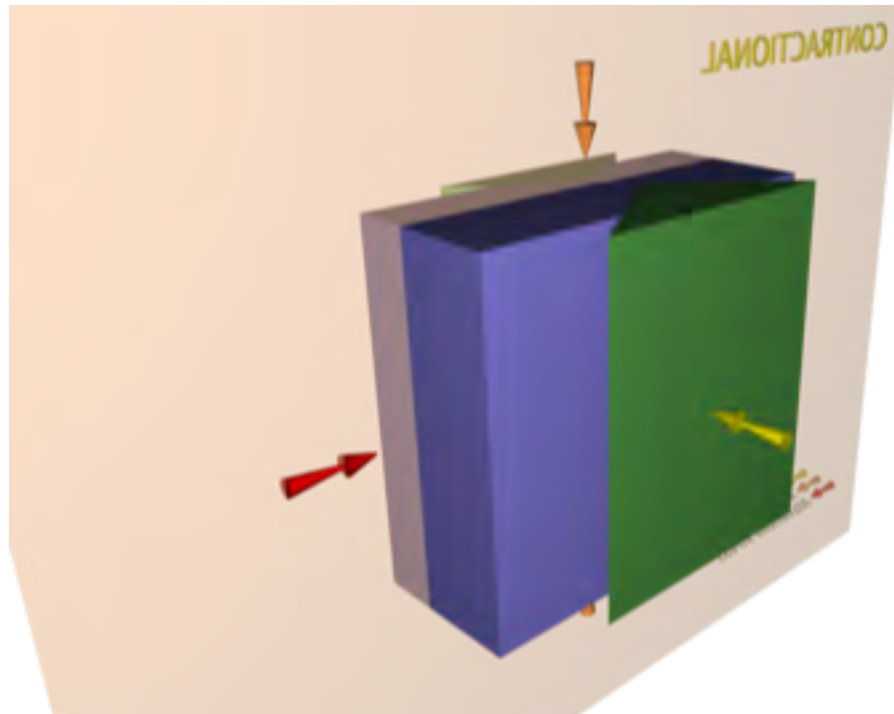
Contractional Tectonic Regime

Extensional Tectonic Regime
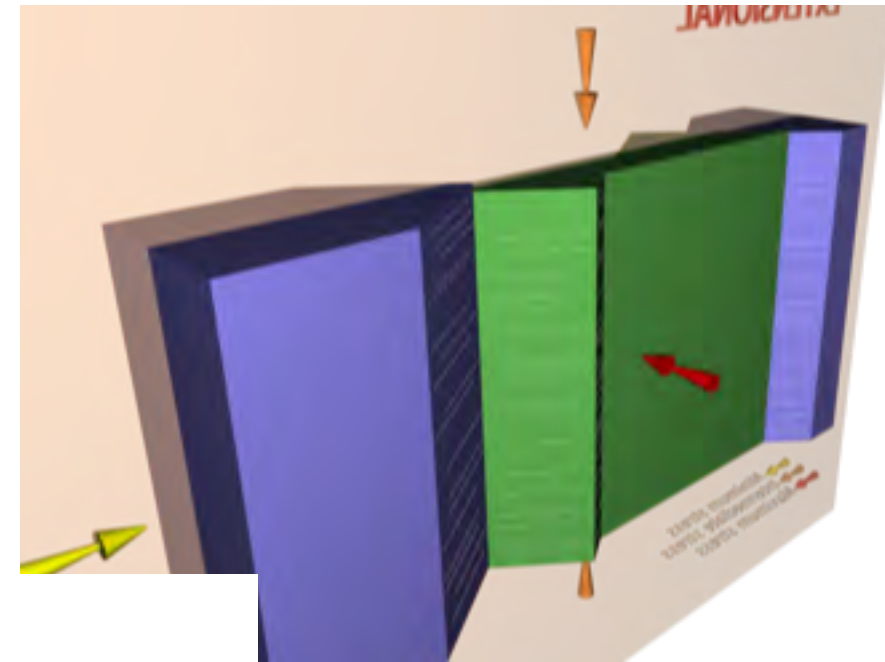
Transcurrent Tectonic Regime
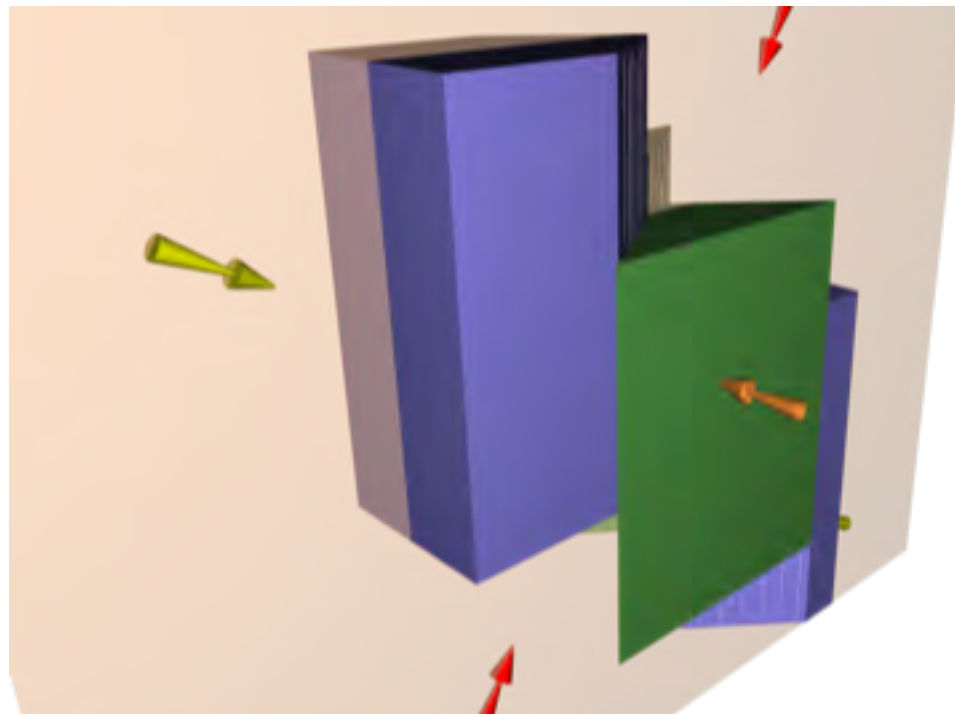
Folds and Folds Systems
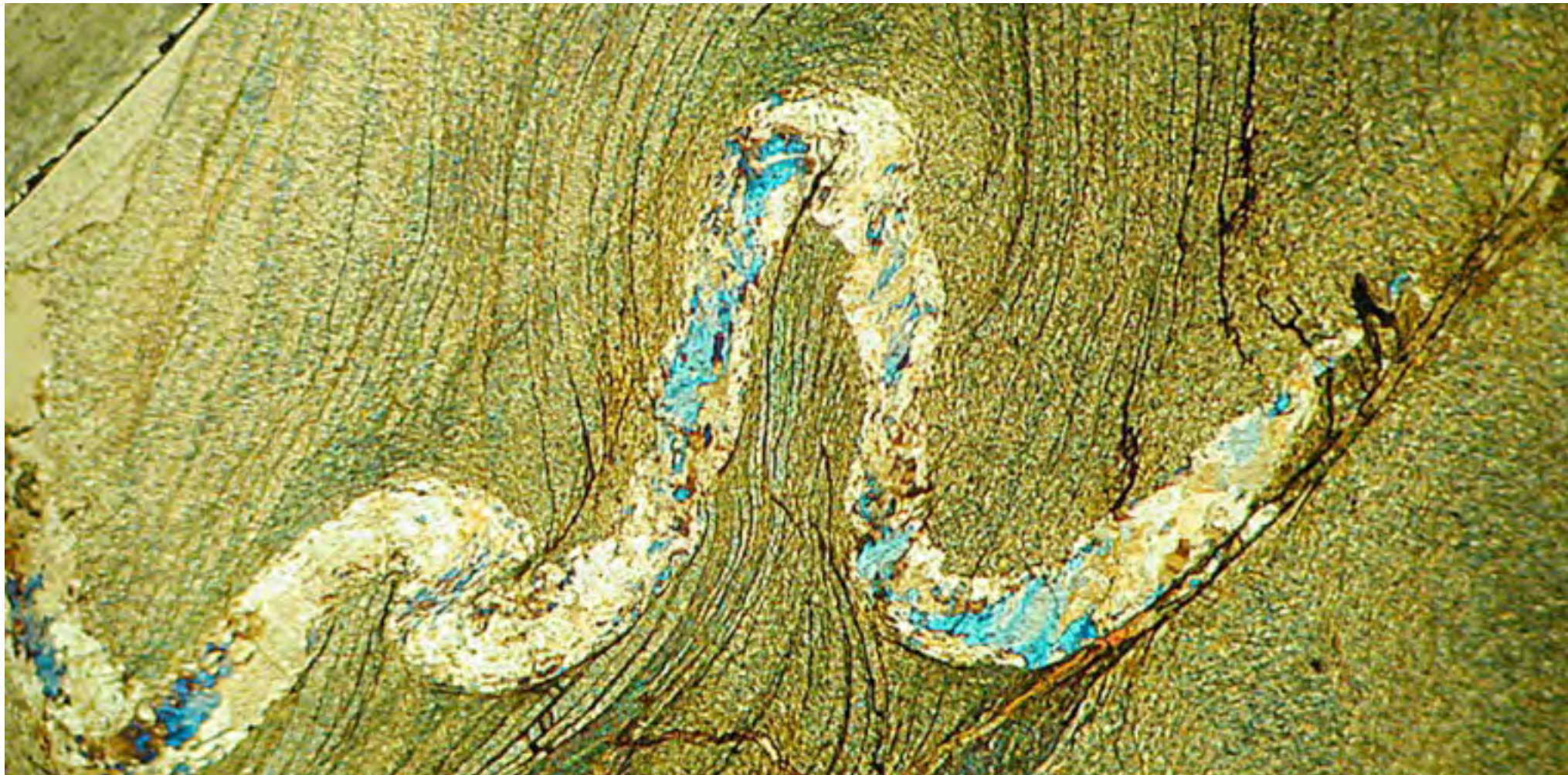
At depth, fracturing and faulting are inhibited by increasing confining pressure as well as increasing temperature which give rocks the ability to flow rather than break. At depth, rock formations organized into layered systems of alternating stronger and less strong layers, buckle under the action of deviatoric stress to form folds and fold systems, or alternatively stretch to form boudins.

Folds Descriptors

Folds form when shortening affects a layered rock formation involving alternating competent and less competent layers. Upon folding, rock formations are bended and buckled into a series of **antiform** and **synform** folds. A **hinge** zone is a region where the dip of the folded surface changes over a small distance. Such a region is also called a **fold closure**. In contrast, the dip is rather constant along fold **limbs**. The terms **anticlinal** and **synclinal** apply when an upward younging arrangement of the rock formations has been verified.

The attitude of folds (their orientation in space) is characterized by the attitude of both their **axial surface** and their B axis also called the **fold axis**. The B axis corresponds to the fold hinge line, the line that links points of maximum curvature on a folded surface. The axial surface is defined as the surface containing all the hinge lines on successive surfaces folded about the same fold. The axial surface can be planar of not, and fold axis can be curved.
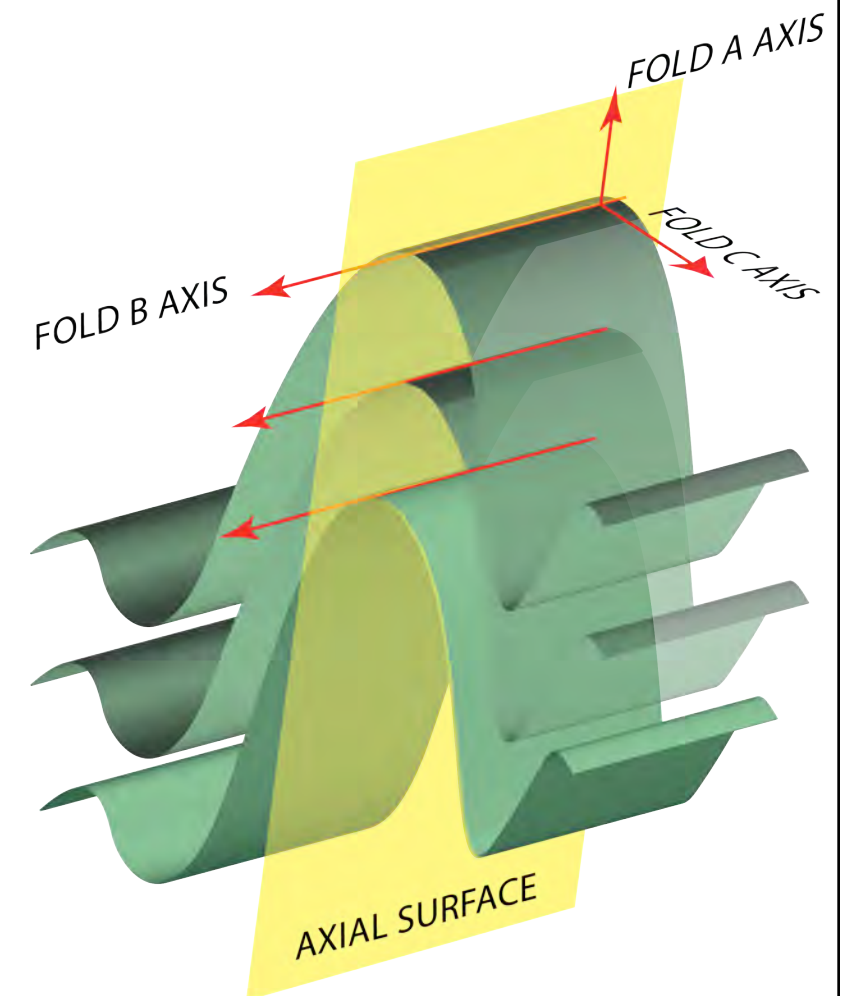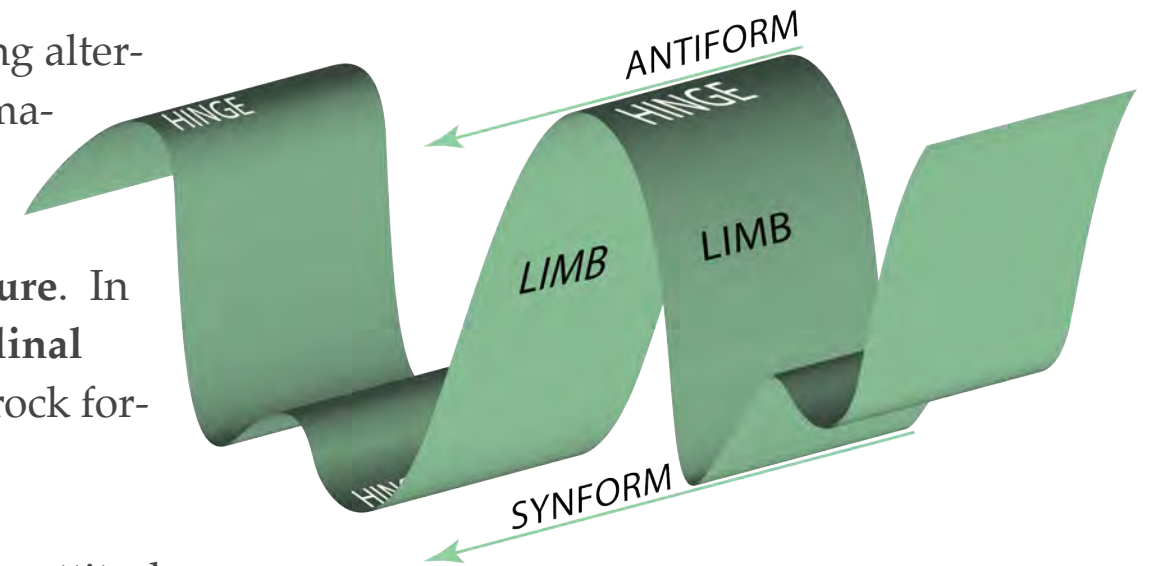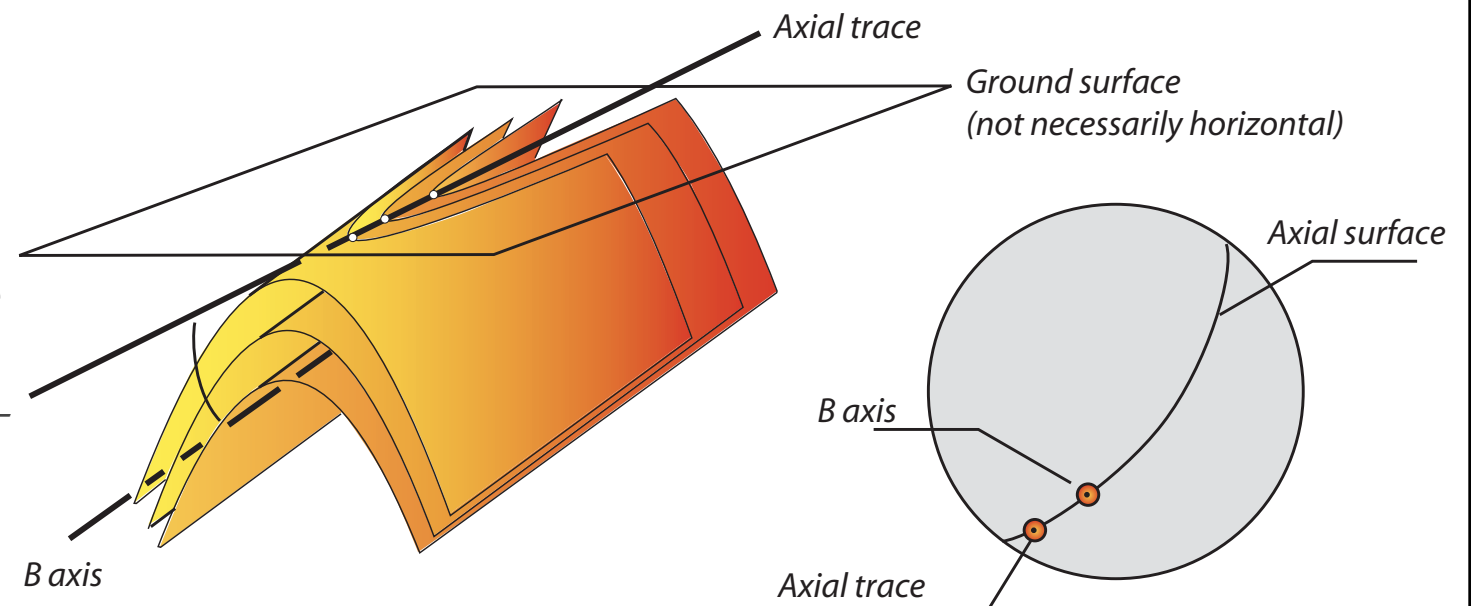
Folds Descriptions Using The Stereonet

Geologists have designed various methods to characterise the geometry of folds from field data, in particular when the size of folds is beyond what can be practically measured directly in the field, typically when fold hinges extend over several hundred meters.

One of these methods involves the use of the stereonet. Before folding, rock formations are sub-horizontal and the poles to beddings are everywhere vertical, hence plotting at the center of the stereonet. However after folding, the poles to the folded surface are no longer vertical but become distributed along a great circle (called the π circle), the pole of which corresponds to the fold B axis. This is know as the π diagram. Alternatively the fold B axis can be inferred from the intersection, on a stereonet, of great circles (β circle) representing the stereographic projection of bedding surfaces randomly distributed about the fold. This is know as the β diagram.
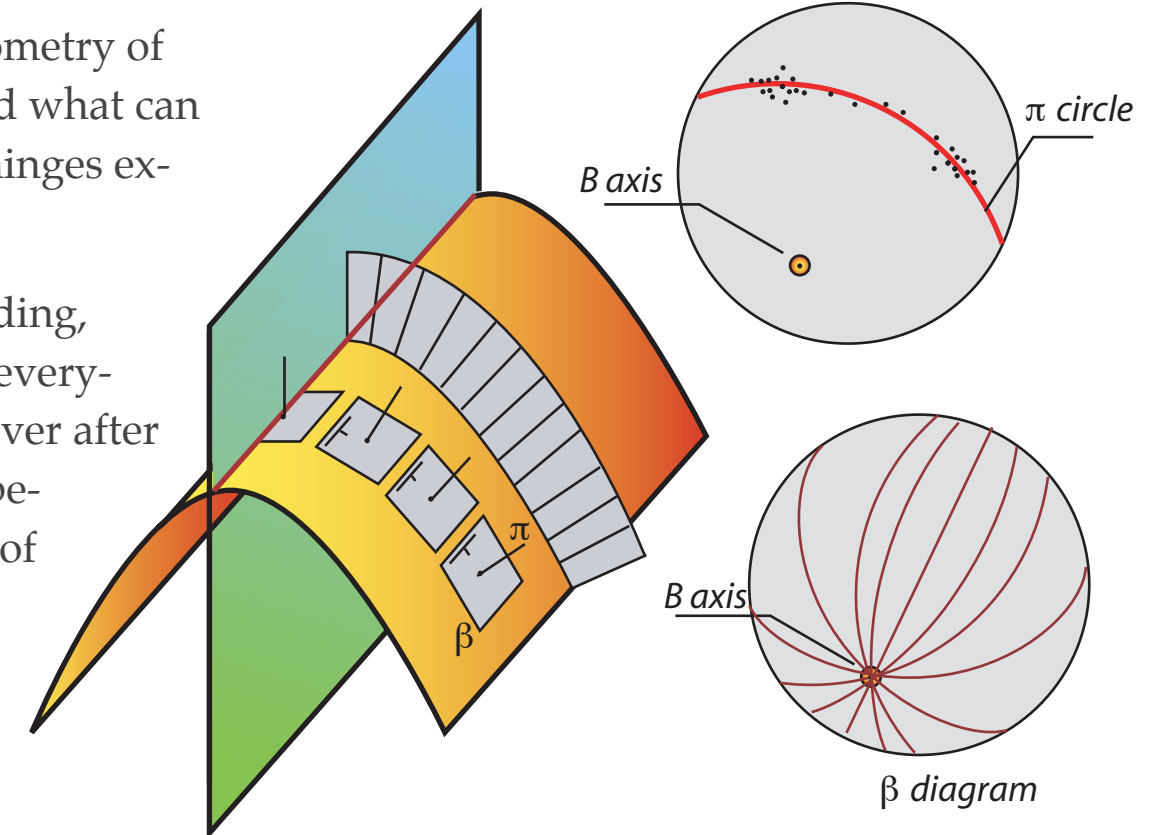
π and β diagrams alone cannot constrain the attitude of the axial surface. This can be achieved by plotting on a stereogram the axial trace of the fold as well as its axis. The axial trace of a fold is the line that links, on the ground surface, points of maximum curvature on successive folded surfaces. By definition the axial trace, since it is defined by points of maximum curvature, lies within the axial surface. Since a planar surface is fully defined by two lines within it (cf. Euclidian geometry), the axial surface can be deduced from the axial trace and the fold hinge.

π diagram, π circle, B axis

β diagram, B axis

Axial trace, Ground surface (not necessarily horizontal), Axial surface, B axis, Axial trace, B axis

Processing pole to bedding data on a stereonet can be used to infer the shape of folds. For instance, π diagrams of isoclinal folds (i.e. folds with parallel limbs) have two strong pole clusters, as random bedding measurements will over-sample limbs compared to that of hinge regions. Asymmetric folds typically have a long limb and a short limb. In this case the long limb will be over-represented leading to one strong cluster and a much smaller one. This contrasts with concentric folds for which pole to beddings are homogeneously distributed over a great circle.

Concentric fold

Isoclinal folds

π diagram

Asymmetric fold
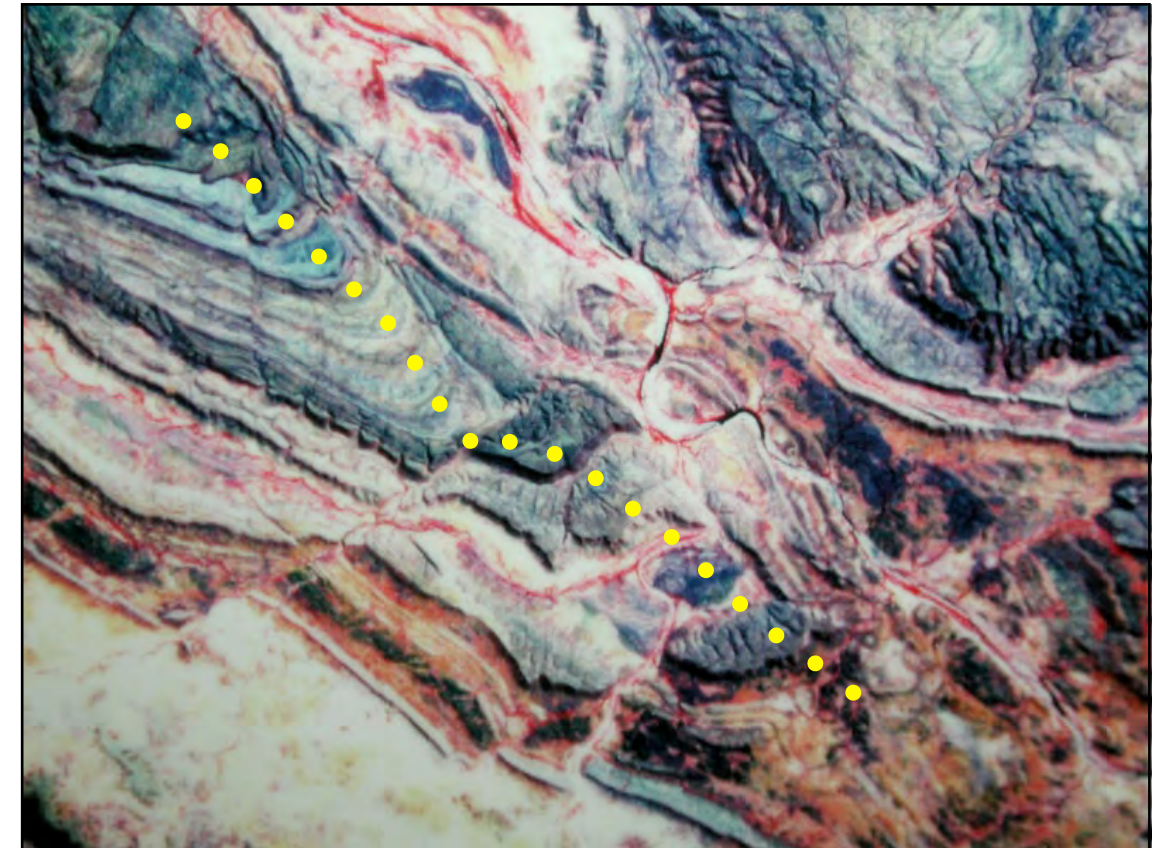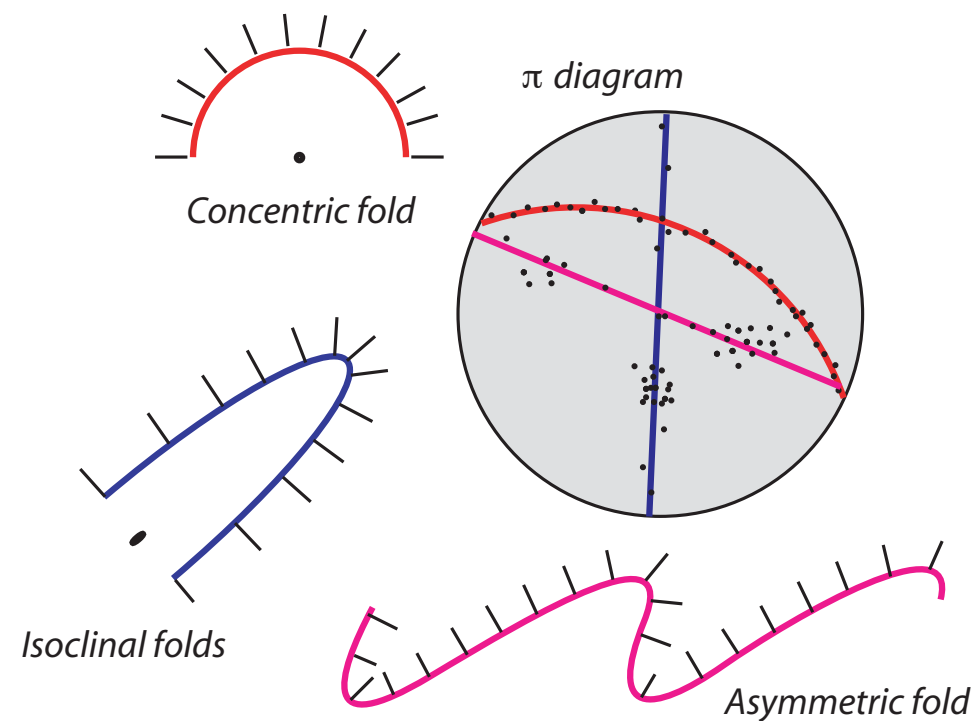
Top: Satellite view of folded Proterozoic rock formations from the Kimberley (WA). The doted line shows the fold axial trace.
Bottom: Upright fold in Carboniferous rocks with Permian unconformity (Portugal, Photo O. Matte).

Folds Classification

There are many ways to classify folds based on: i/ the thickness of folded layers, ii/ the angle between limbs, iii/ the dip of the axial surface, iv/ the plunge of the fold axis, v/ the general shape of folds ... The following introduces a number of terms commonly used to describe folds.

Isopach folds are folds in which the thickness t of each layer keeps constant. In **similar folds** it is the apparent thickness e measured parallel to the axial surface that remains constant.
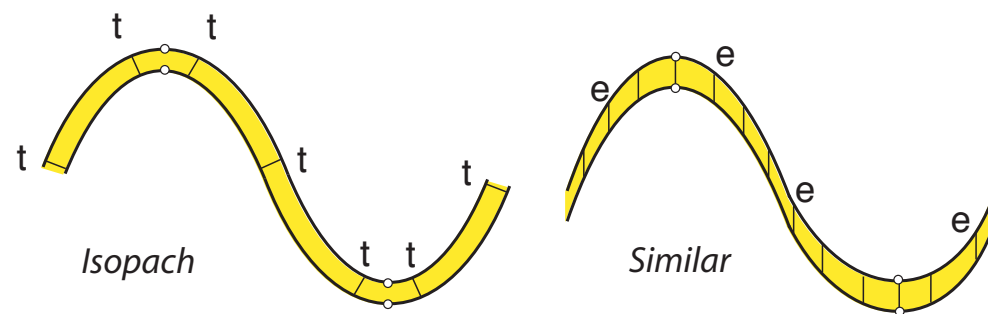
**Open, tight and isoclinal**, are terms referring to the angle between two successive fold limbs.
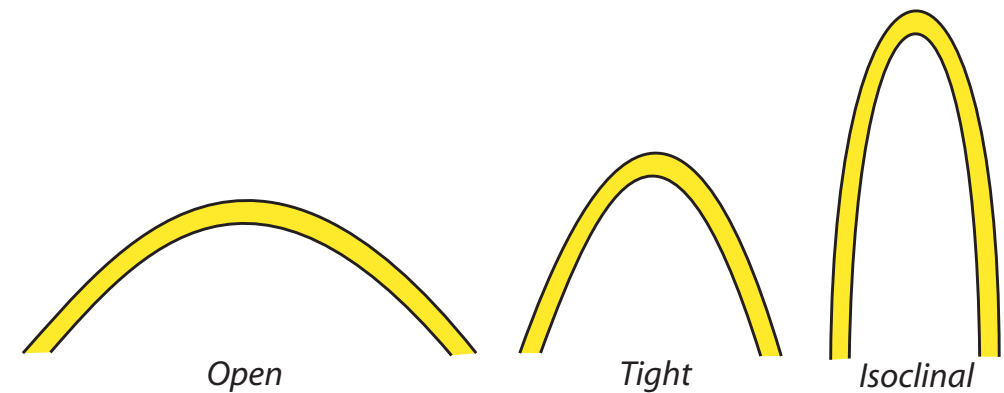
Upright folds have a vertical axial surface. They may or may not be symmetric. **Overturned folds** have one limb with the younging direction pointing downward, whereas **recumbent fold** have sub-horizontal axial surfaces.
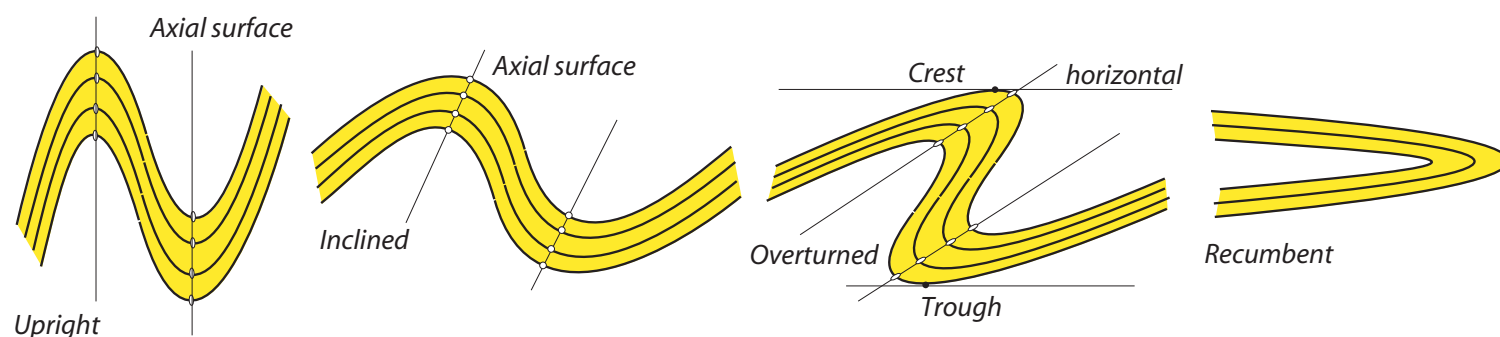
Recumbent folds in gneiss beneath the Indus Suture (Photo Credit: Gerold Zeillinger, diogenes.ethz.ch)

Horizontal, **vertical**, **plunging** and **reclined** are terms used to describe the plunge of the fold axis:
•Horizontal folds have horizontal fold axes.
•Vertical folds have vertical fold axes.
•Plunging folds have plunging fold axes.
•Reclined folds have plunging fold axes AND younging pointing downward.

Plunging fold

Vertical fold

Reclined fold
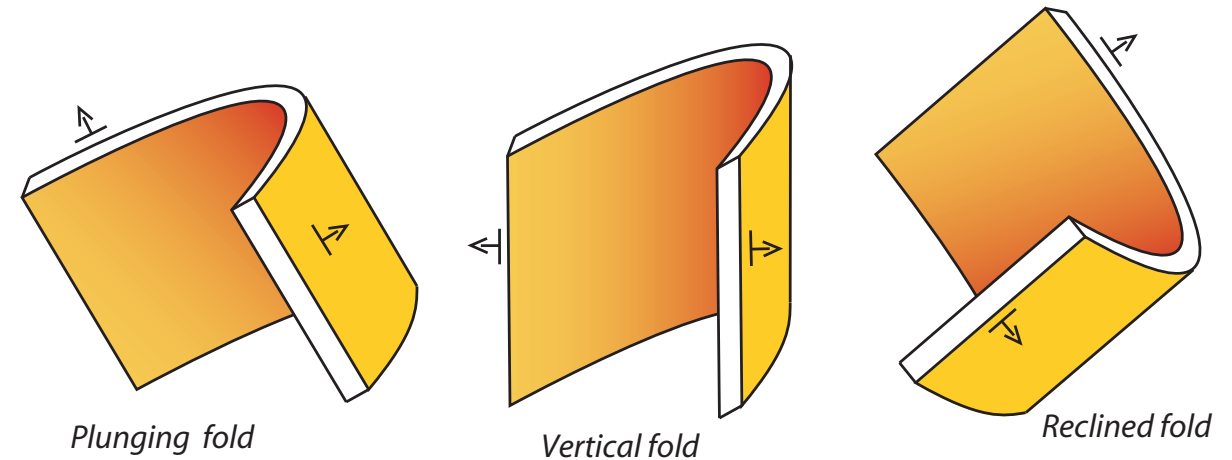
Folds can have all sorts of complex shapes and consequently all sorts of weird names.
•**Concentric** folds have a center of symmetry.
•**Chevron** folds (also called **kink** folds) have narrow hinges relative to the length of their limbs.
•**Ptymatic** folds have short limbs relative to the their hinges.
•**Box** folds somehow have three limbs and two hinges.
•**Rootless** folds have broken limbs due to shearing.

Gallery 4.1 Folds

Concentric fold

Kink fold

Ptygmatic fold

Box fold

Rootless

Chevron folds in a paragneiss (Olary block, SA, Photo P. Rey).
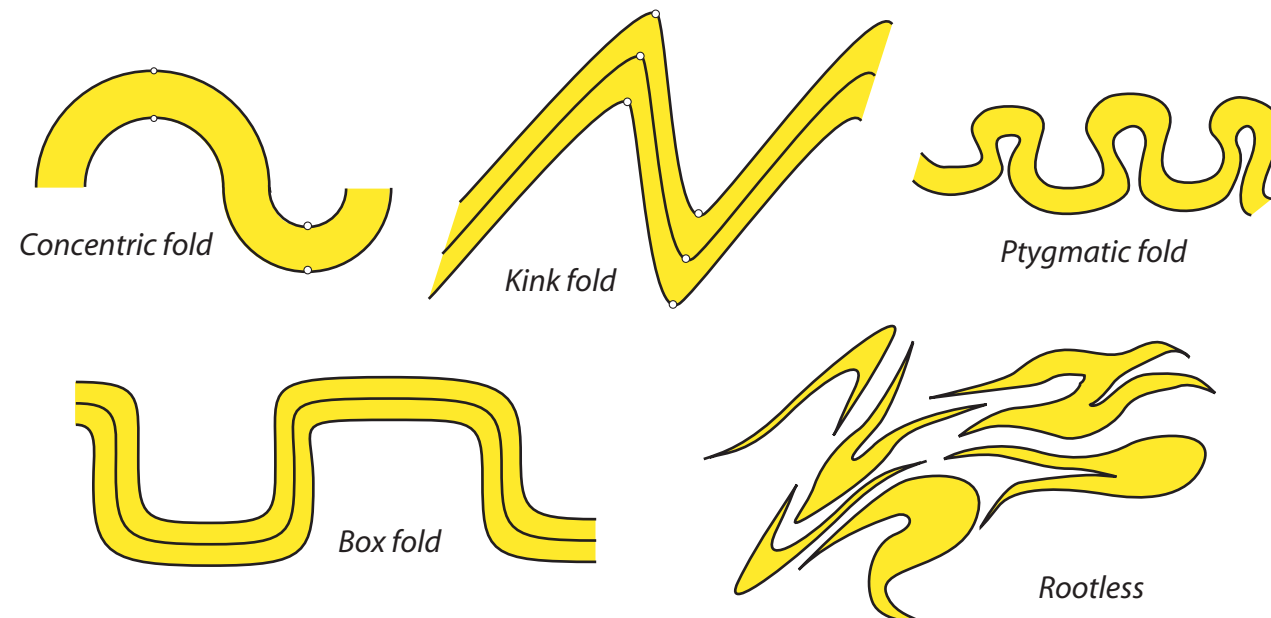
Fold Related Microstructures

During folding, the bending of bedding surfaces (S_0) is accommodated by the deformation of rocks at the grain scale. This deformation may produce the rotation and alignment of mineral grains and biogenic clasts parallel to the fold axial surface. This produces a secondary anisotropy plane (i.e. a planar fabric) called a cleavage plane (S_1).

Deformation can also change the shape of grains through a mechanism called pressure-solution. According to this mechanism of deformation, the part of the grain enjoying the larger stress is dissolved and the material removed migrates towards areas of lower stress. Through pressure-solution, a grain of spherical shape is transformed into an ellipsoid whose flattening plane (defined by its long and intermediate axes) tends to be parallel to the fold axial surface, hence participating in the development of a cleavage plane.

Two other mechanisms contribute to the development of axial planar cleavage. Fracture cleavage involves the development of shear fractures in competent (hard) layers. These shear fractures tend to be parallel to the fold axial surface. Finally, crenulation cleavage develops because of the preferential alignment of limbs in micro-folds.

Crenulation cleavage

S_1

S_0

Parasitic fold

Fracture cleavage

Slaty cleavage

S_0: Bedding plane
S_1: Axial plane cleavage

1 cm - 1km

Crenulation in a low-grade schist, the photo is ca. 2cm across. On the right small detail of the above photo (Photo P. Rey).
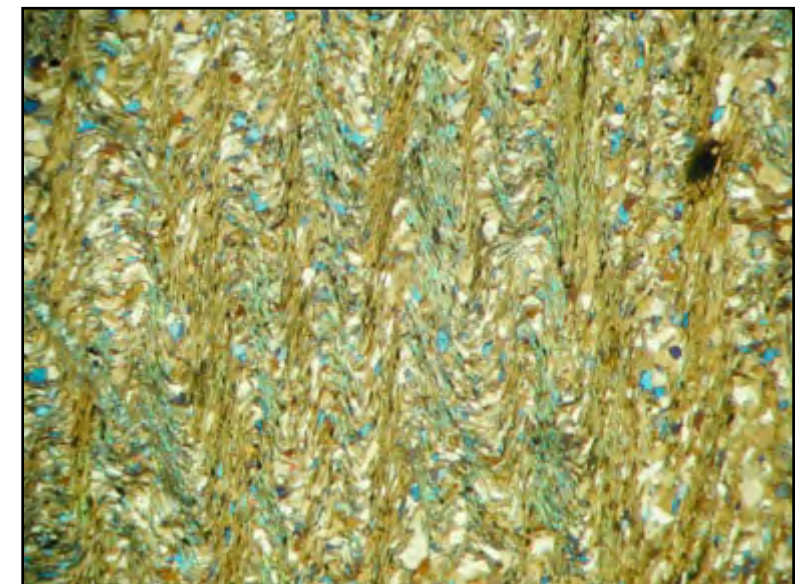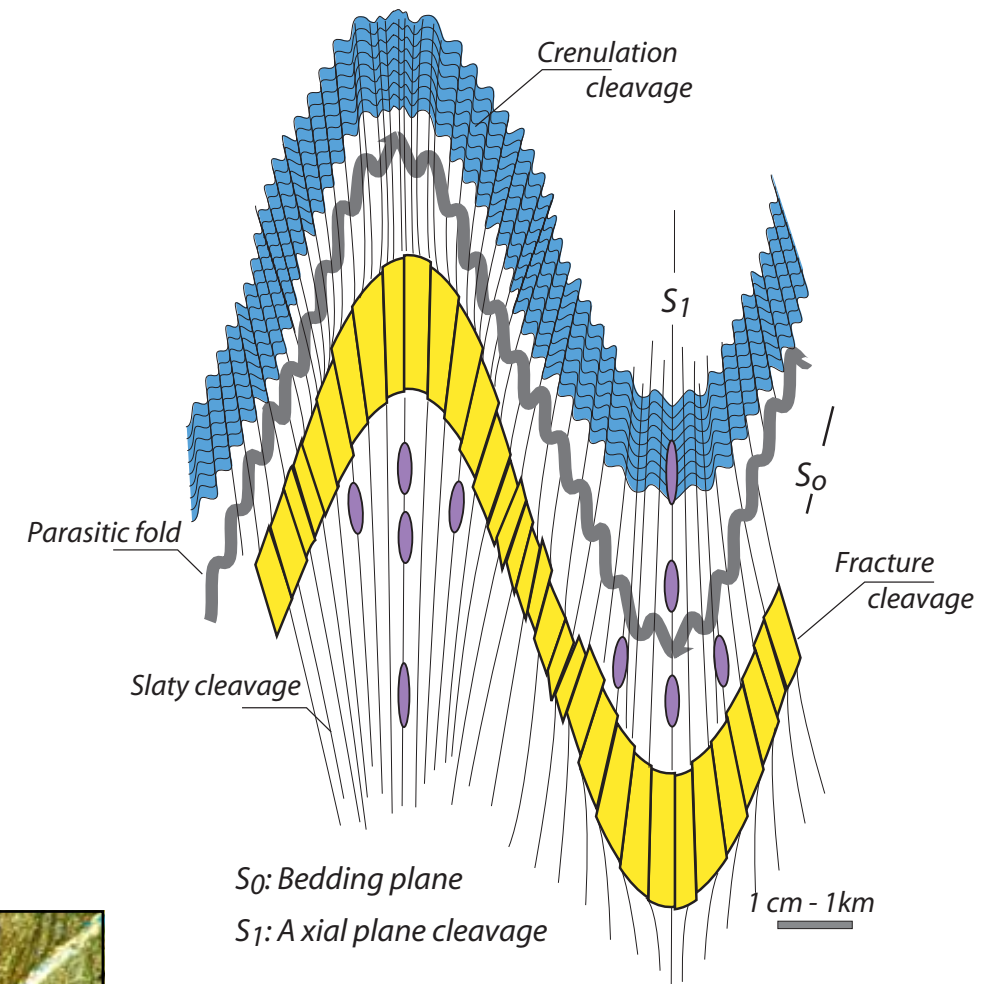
Linear structural features associated with folding develop as the result of micro-folding, and intersection between S_0 and S_1 planes.

Fold rodding lineations, also called **crenulation lineations**, correspond to the hinge of micro-folds along bedding planes. Cleavage-bedding intersections are called **intersection lineations**. These lineations appear on cleavage and bedding surfaces.

Both types of lineation run parallel to the fold axis, and therefore can be used as a proxy for fold hinges (i.e., the plunge and plunge direction of these lineations is the same as that of the fold axis).
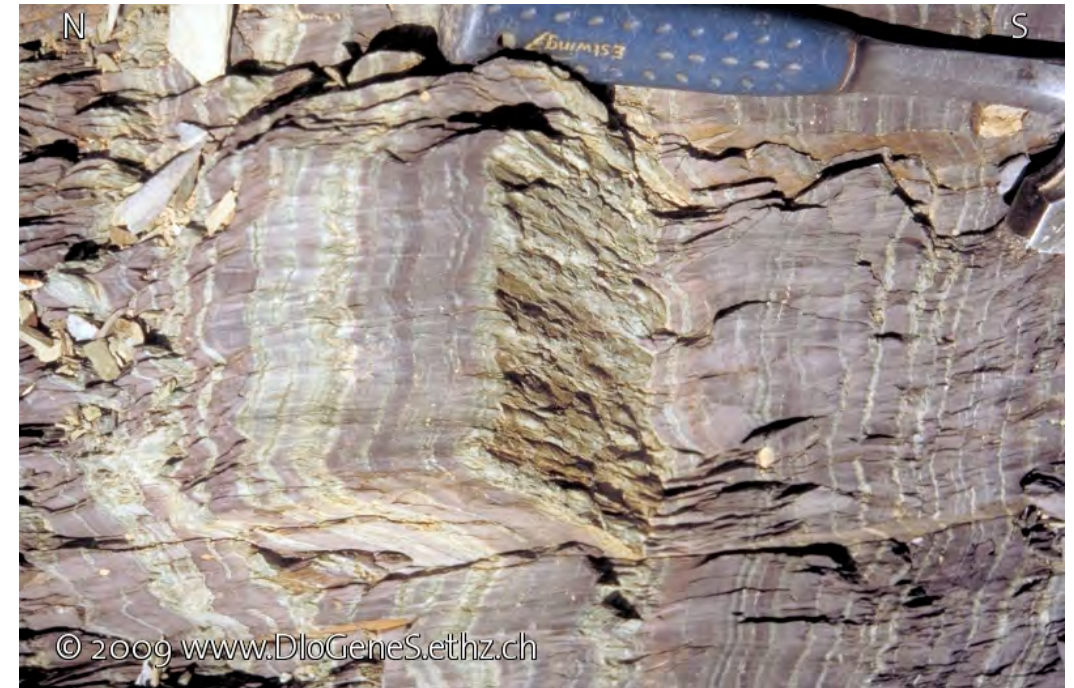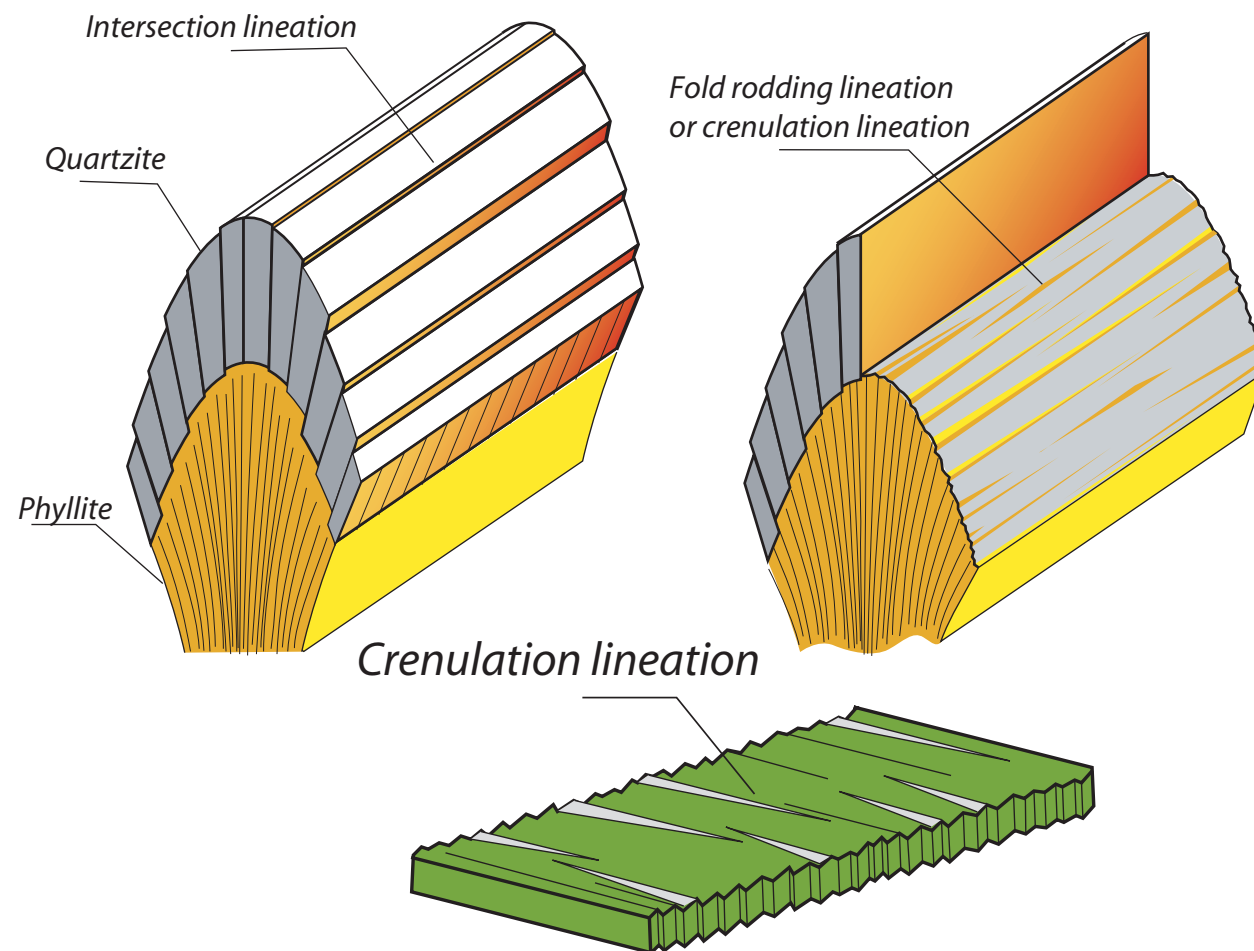
Bedding and cleavage fabrics in a low grade schist (Photo credit: J.-P. Burg, diogenes.ethz.ch)

Intersection lineation

Fold rodding lineation or crenulation lineation

Quartzite

Phyllite

Crenulation lineation

Crenulation lineation in a low grade schist (Photo credit: J.-P. Burg, diogenes.ethz.ch)

Fold Systems: Cleavage-Bedding Relationships

One problem that geologists face is figuring out, in poor outcropping areas, the succession of anticlinal and synclinal closures adjacent to a given outcrop. The geologist below tries to figure out the position of the next anticline and syncline closures with respect to the outcrop. The outcrop consists of four beds dipping to the west. There is a cleavage plane (S_1) also dipping to the west, but much steeper that the bedding (S_0).

The sketches A1 and B1 shows two possible solutions: either the adjacent anticlinal closure is to the left (A1), or it is to the right (B1) of the outcrop. In the next outcrop to the right (sketches A2 and B2), the geologist observes that the succession of rock formation is in the reverse order suggesting that a fold closure has been passed. To find out which of A2 or B2 fold systems is correct, we must remember that cleavages diverge away towards fold closures. From the angular relationships between S_0 and S_1 there can be only one valid solution which is given on the next page...
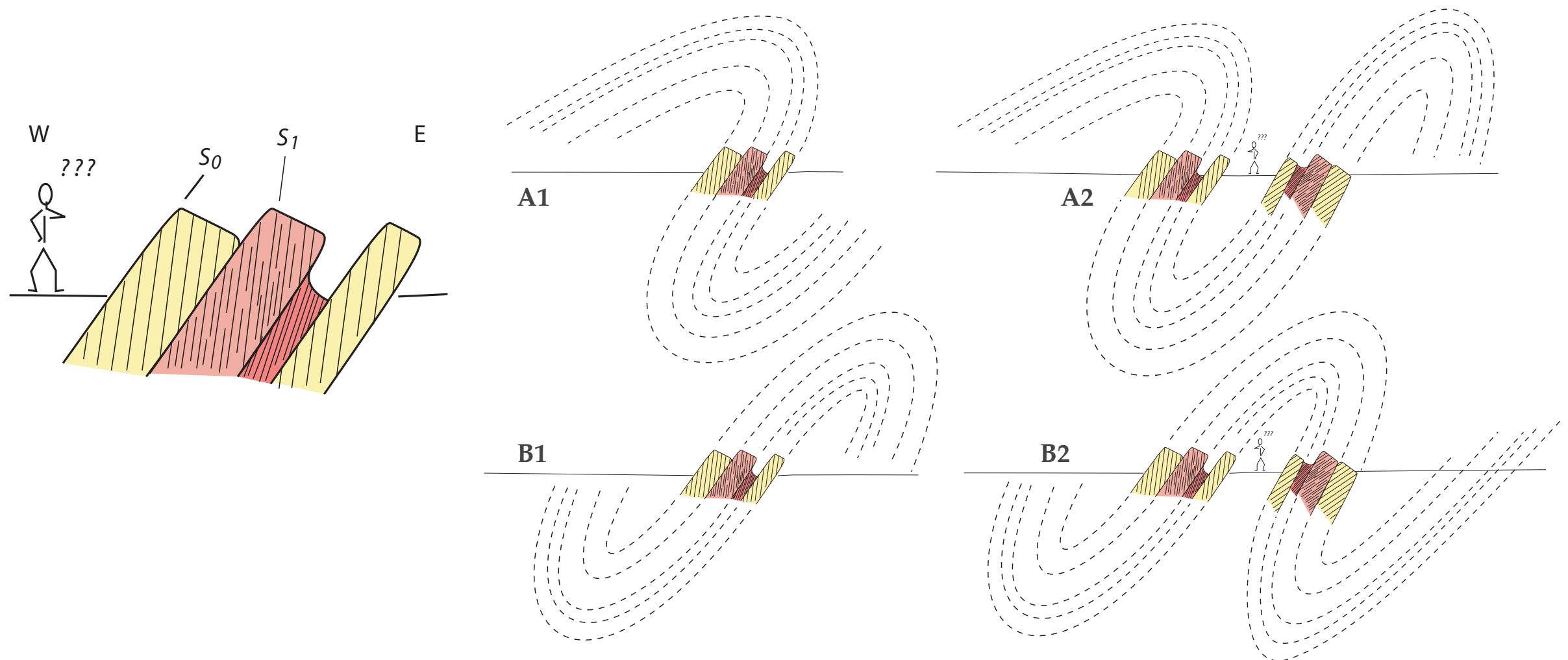
Cleavage diverges away towards the fold closure. The correct solution is therefore given by the sketch B2, where the cleavage diverges upwards away from the core of the anticline. In contrast the sketch A2 shows that the cleavage converges downwards toward the core of the synform, a situation which is not observed in nature.

In conclusion, cleavage-bedding relationships help structural geologists 1/ to infer the position of a given outcrop in a fold system, 2/ to infer the succession of adjacent fold closures with respect to an outcrop.
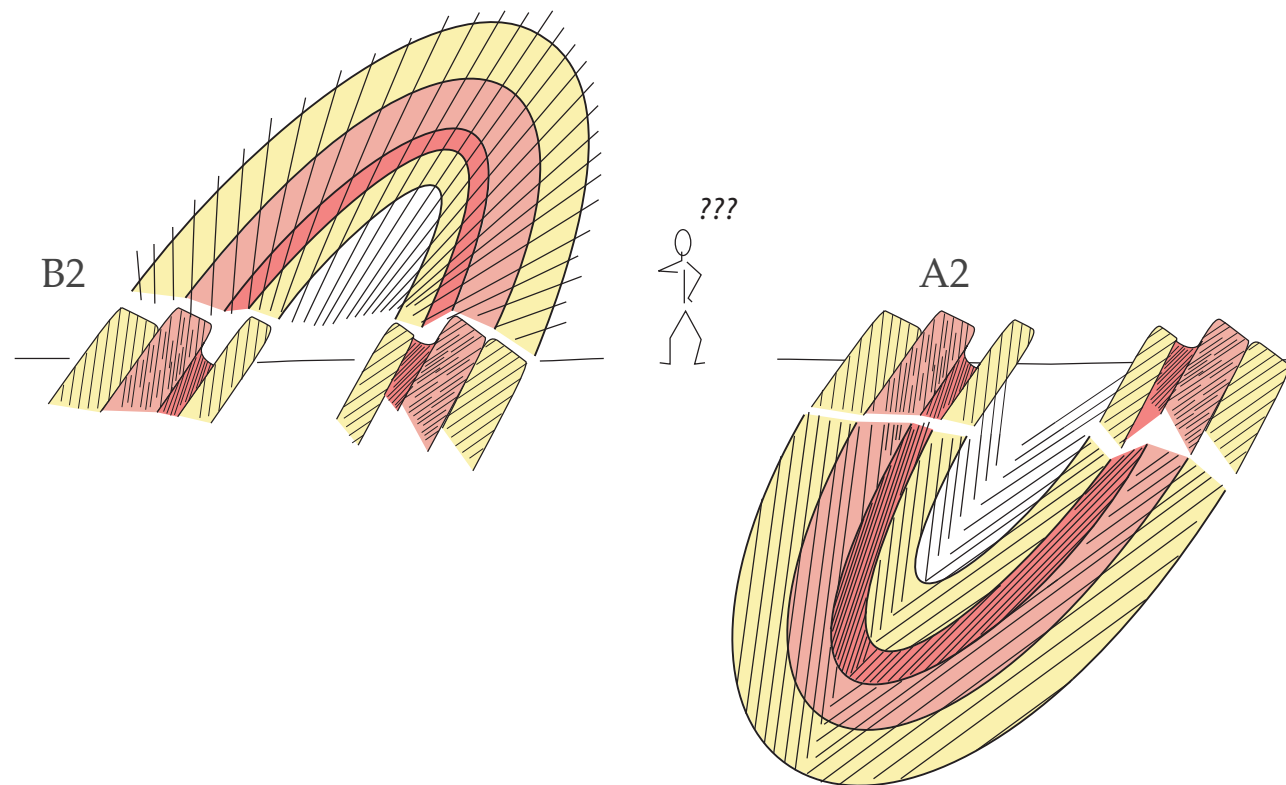
Cleavage may not always develop with a fan distribution. Here an axial planar cleavage parallel to the fold axial plane in low grade schist. This suggests that there is little strength contrast between beds (Photo credit: P. Rey). NB: Stratigraphic layering is shown by change in color, texture and/or weathering.

Fold Systems: The Concepts Of Vergence & Facing

The **facing** of a fold system refers to the geographic direction of younging (shown with an arrow) of the long limbs of its parasitic folds. In a fold system, such as the one presented below, the facing is changing as one crosses major fold hinges. Indeed from west to east, this fold system is facing west upward, then east downward, then west upward, and east.

The concept of **vergence** refers to the general sense of shear involved in the development of asymmetric folds. The fold system shown below is verging east as the asymmetry is the result of top toward the east shearing. Unfortunately vergence is also used to refer to the direction of the next anticlinal closure.

Changes in fold asymmetry, cleavage-bedding relationships, facing, and shear vergence occurs across fold hinges. Determination of two of these criteria constrain the two others.


Younging criteria are sedimentological features informing about the relative position between younger and older layers.
Left: Loading cast sedimentary structure showing upward younging.
Right: Example of cross-beds showing younging to the left (Photo Credit: P. Rey).
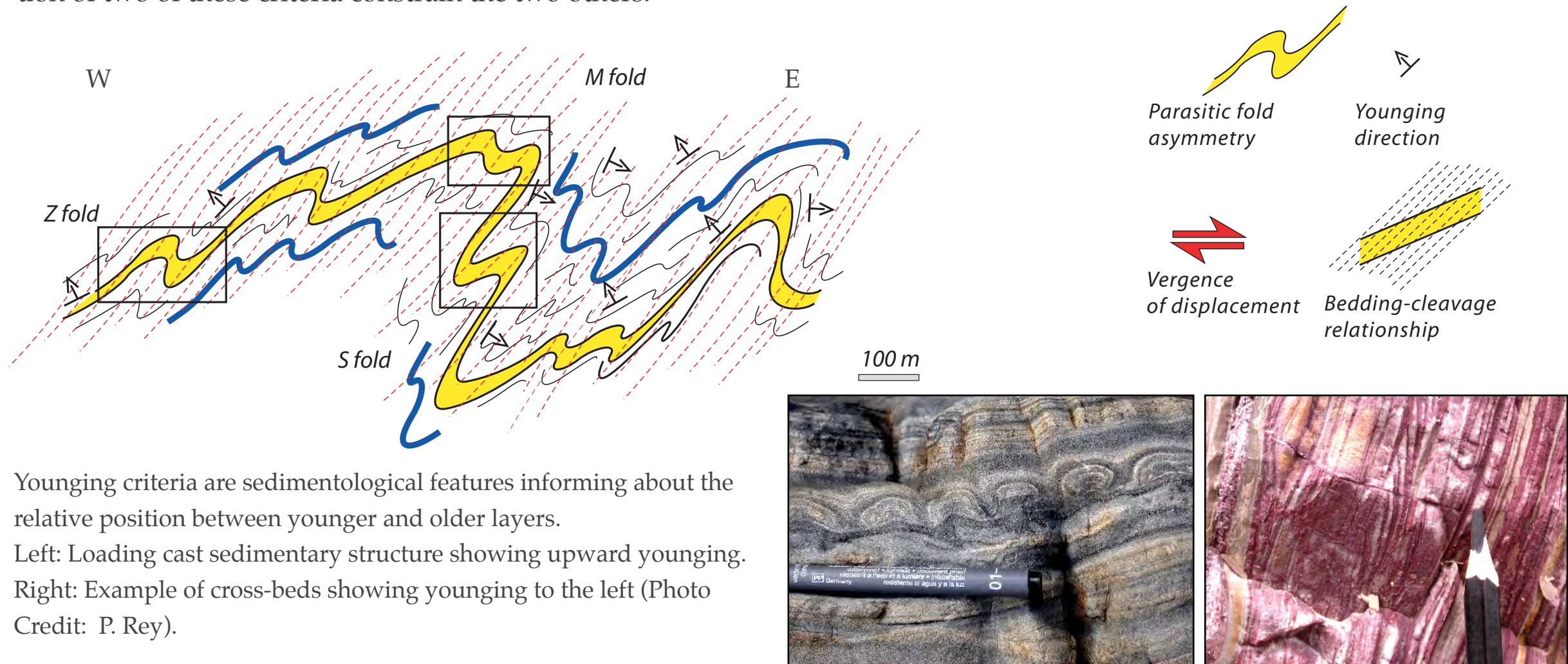
Mind The Gap …

The asymmetry of parasitic folds involves either a clockwise rotation (Z-shaped folds) or an anti-clockwise rotation (S-shaped folds) of the folded surface. The asymmetry of parasitic folds changes across fold hinges as S-shape folds occur in one limb and Z-shaped folds occur on the other. Therefore one can use this change to track down the position of large-scale fold hinges in regions of poor exposures.

When doing this some caution should be exercised. In particular it is important to realize that the sense of fold asymmetry (S or Z) depends on the direction toward which the observation is made. The sketch on the right shows a parasitic fold. Looking from east to west the fold appears as a Z-shaped asymmetric fold. In contrast, when looking from west to east the fold appears as an S-shaped fold. This change in asymmetry is not the result of having crossed a fold hinge but simply the consequence of looking at the same parasitic fold in two different directions.

To avoid any mistake, fold asymmetry should be determined looking toward the same direction, usually toward the plunge direction of the fold axis in plunging or reclined folds.

Another very common source of mistake is entrenched in the fact the 2D sections through 3D objects can be extremely misleading in representing the true fold morphology. Indeed, a simple upright symmetric fold can appear as an apparent overturned to recumbent fold on a 2D vertical section cutting at an angle to the fold axis.

W   E
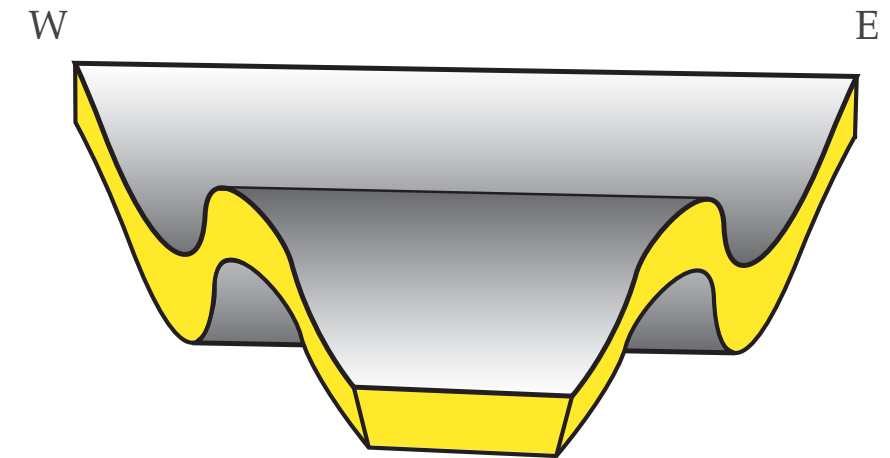
Movie 4.1 Mind the gap!
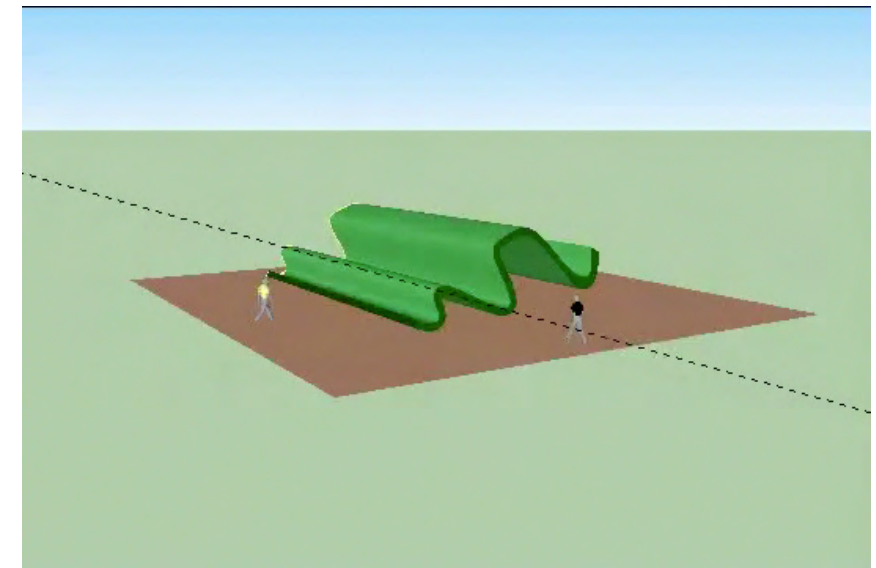
A Little Quiz ...

1/ What is the fold axis?

2/ What is the axial trace?

3/ What is the axial planar surface?

4/ Draw an overturned, tight, isopach fold.

5/ Draw an inclined, open, similar fold.

6/ What is a recumbent fold?

7/ What is a reclined fold?

8/ What do we call a fold with long limbs and narrow hinges?

9/ What is a crenulation cleavage?
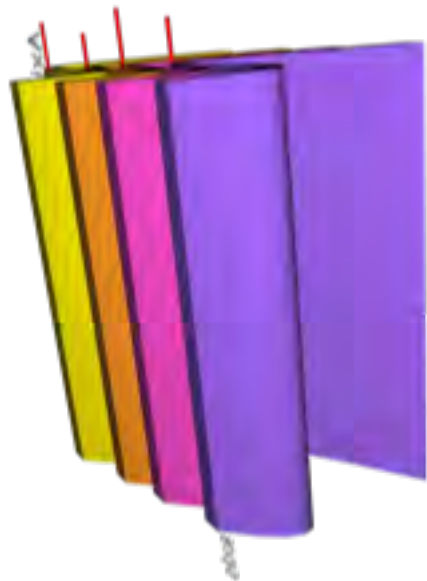
10/ What information do intersection lineations provide?

Comment on the geometry of the folds (Photo Credit: P. Rey).

● ● ● ● ●

Interactive 4.1 Fold axis and fold axial surface.

Question 1 of 6

The attitude of folds and fold systems is or can be fully characterized by:

○ **A.** Strike, dip and dip direction of the axial planar surface.

○ **B.** The pole to beddings plotted on a stereonet canvas.

○ **C.** The plunge and plunge direction of the fold axis.

✓ **D.** A and C together.

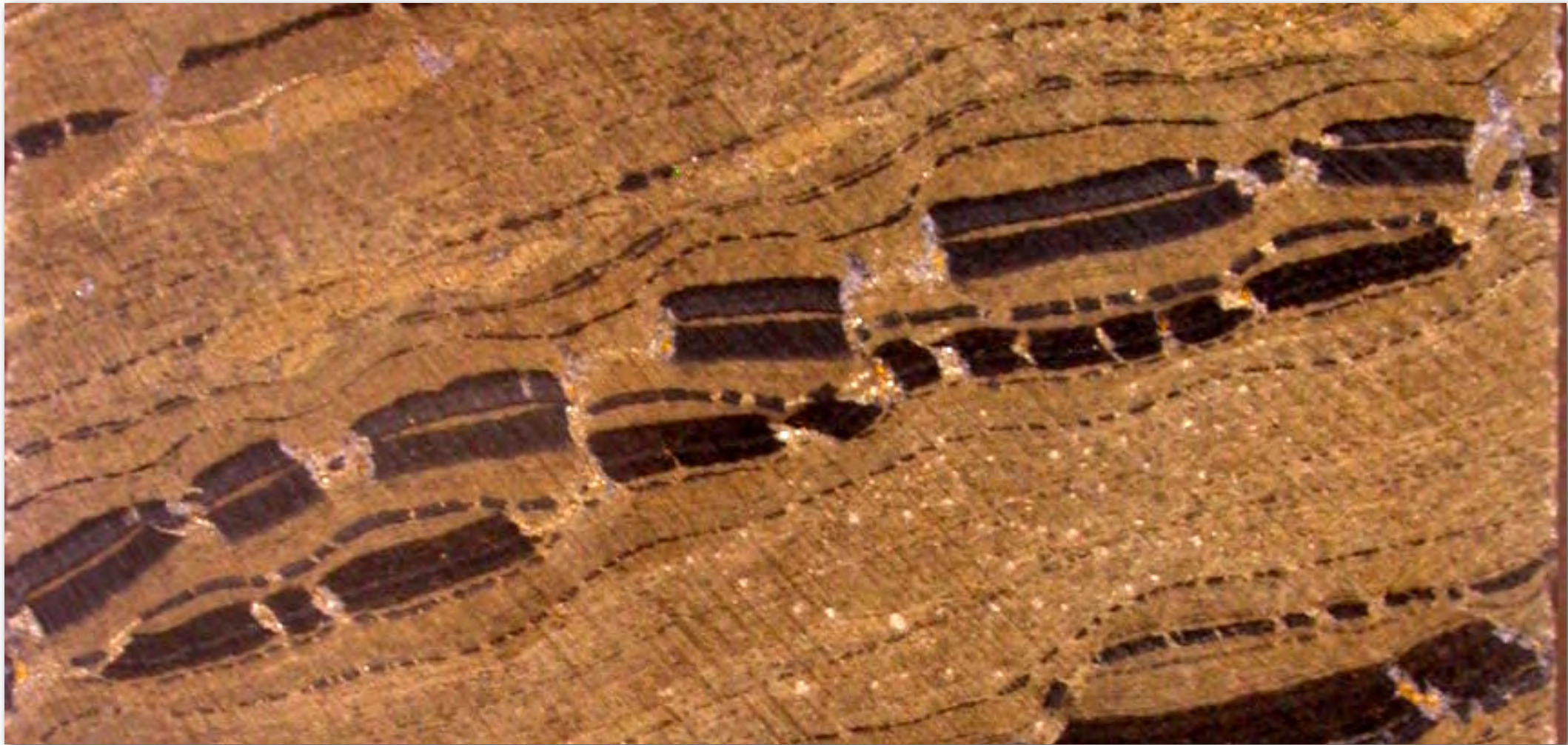✓ **E.** The plunge and plunge direction of the fold axis, and the plunge and plunge direction of an axial trace

◀   **Check Answer**   ▶

CHAPTER 5
Boudins and Boudinage

We have seen in the previous chapter that folds are an expression of contractional deformation. In this chapter we look at boudins and boudinage which are the result of extensional deformation. Like folding, boudinage exists because rock formations are made of alternating stronger and less strong layers. Contraction and extension can be local and independent of the tectonic setting. Indeed, folding and boudinage can develop at the same time in all tectonic regimes.

Boudins Descriptors

Boudins form due to a process called **boudinage**. Boudins form when lengthening affects a layered rock formation involving competent layers embedded into a less competent, easily deformable, host rock.

Upon extension, the stronger layers lengthen via heterogeneous thinning leading to the development of **pinch and swell** structures (i.e. thinning of the strong layer is periodic). In contrast, the weaker layer can stretch and thin homogeneously.

In the stronger layers, amplification of thinning in the pinched regions eventually leads to their segmentation into boudins separated by **necks**.

The strength contrast and the direction of extension with respect to the layer impose a strong control on the geometry of the boudins, which can be either symmetric or asymmetric.

20 cm

Pinch-and-swell structures
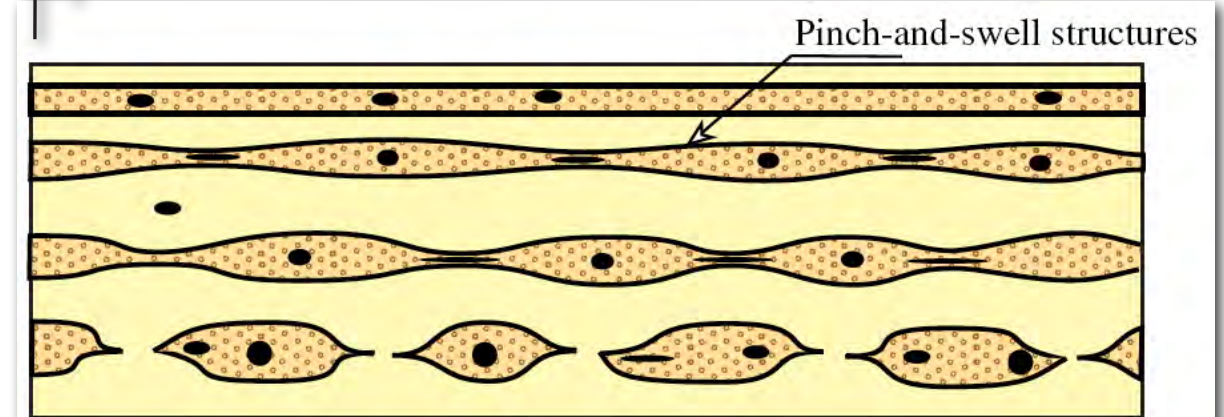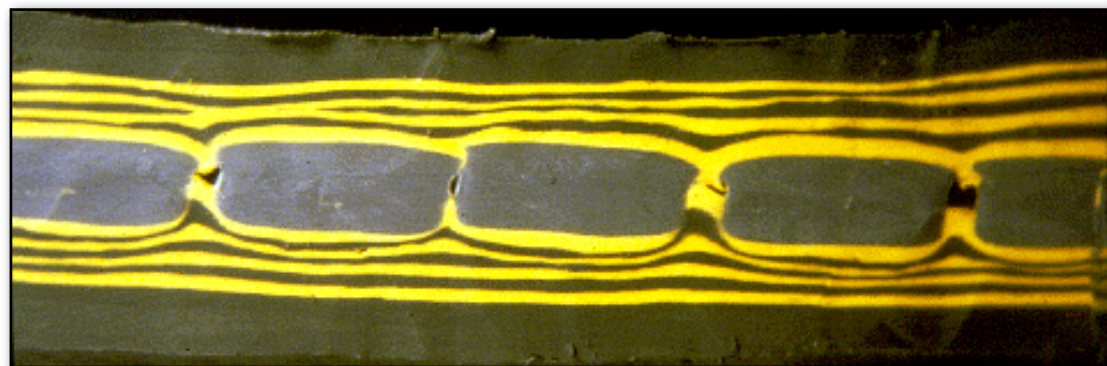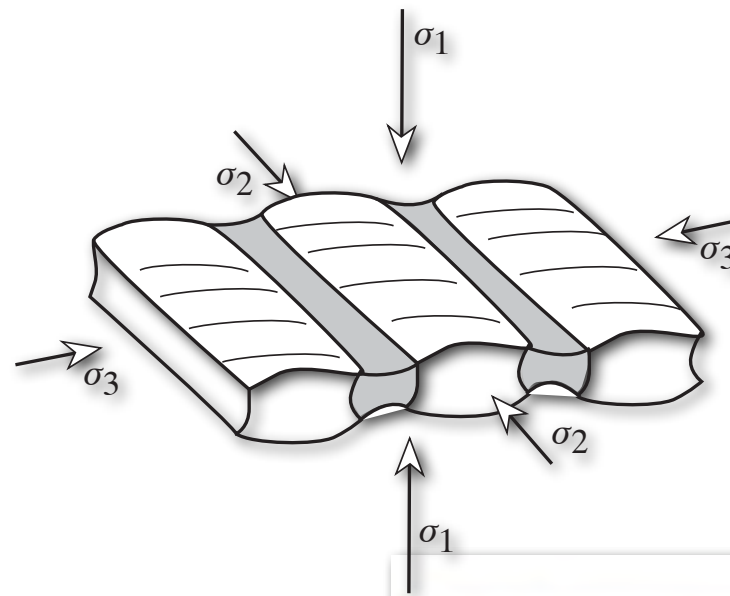
Symmetric Boudinage

The direction of extension with respect to the orientation of the layering imposes a strong control on the geometry of the boudins.

Extension parallel to layering leads to symmetric boudinage. Thinning and necking in the stronger layers often initiates via the development of extensional fractures orthogonal to the layering. As extension proceeds, boudins evolve with a symmetric shape, moving apart in a direction parallel to the original layering.
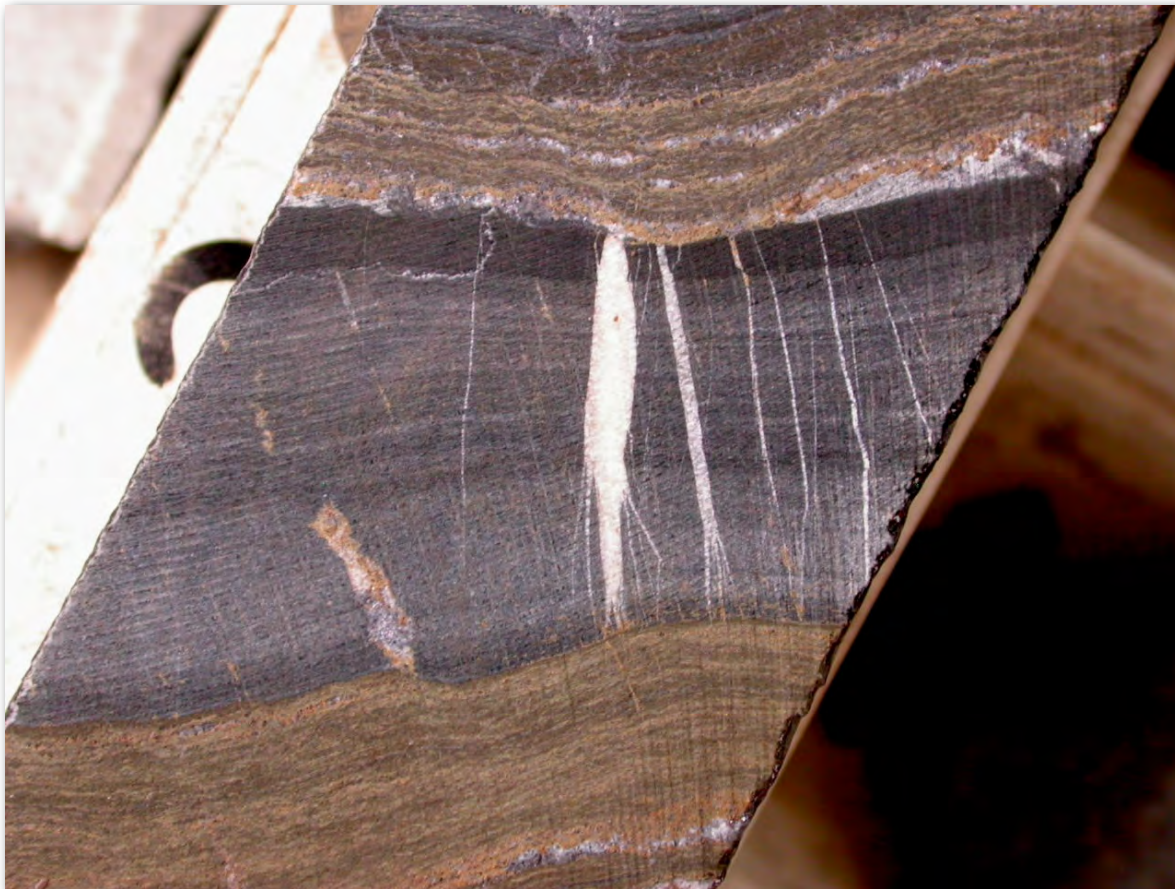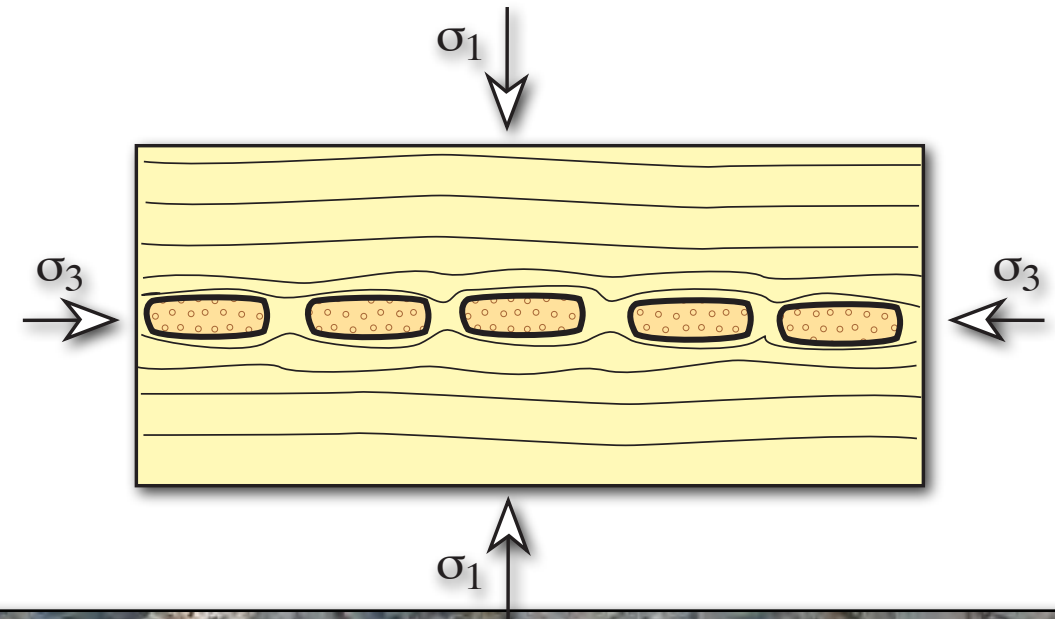
Example of extensional fractures orthogonal to bedding in a nascent neck region, the photo is ca. 8 cm long (Photo credit: P. Rey).

Example of symmetric boudinage of a granitic vein in a psammitic schist (Broken Hill, NSW, Photo credit: P. Rey).

Asymmetric Boudinage

Asymmetric boudinage can develop following two processes.

1/ When extension occurs at an angle to the layering, necking and thinning in stronger layers can initiate via the development of shear fractures or shear bands depending on the temperature. The boudins evolve with a strong asymmetric shape. As extension proceeds boudins rotate to become parallel to the foliation plane, while developing a sigmoid shape.

2/ Asymmetric boudinage can also occur during shearing when the shear zone cuts through a layered rock formation. In this case, the sense of shear can be inferred by the sigmoid shape of the boudins, and/or the curvature of the fabric of the host rock against the shear bands.

Gallery 5.1 Asymmetric boudinage
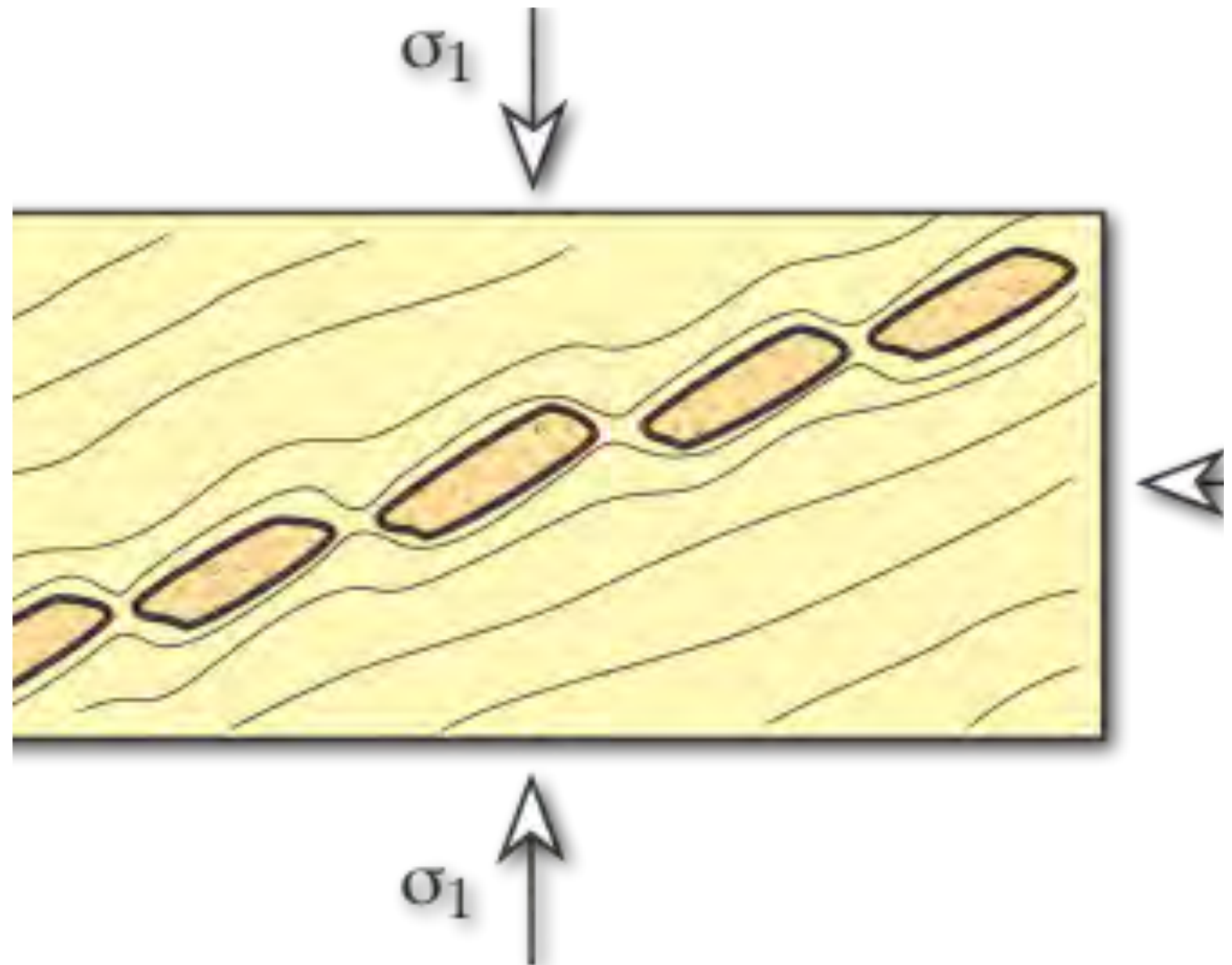
Mind The Gap ...

Folding and boudinage are NOT characteristics of contractional and extensional tectonic regimes. Both types of structure can be found in any tectonic regimes. In fact, boudinage can be the consequence of folding, since fold limbs tend to lengthen during deformation.
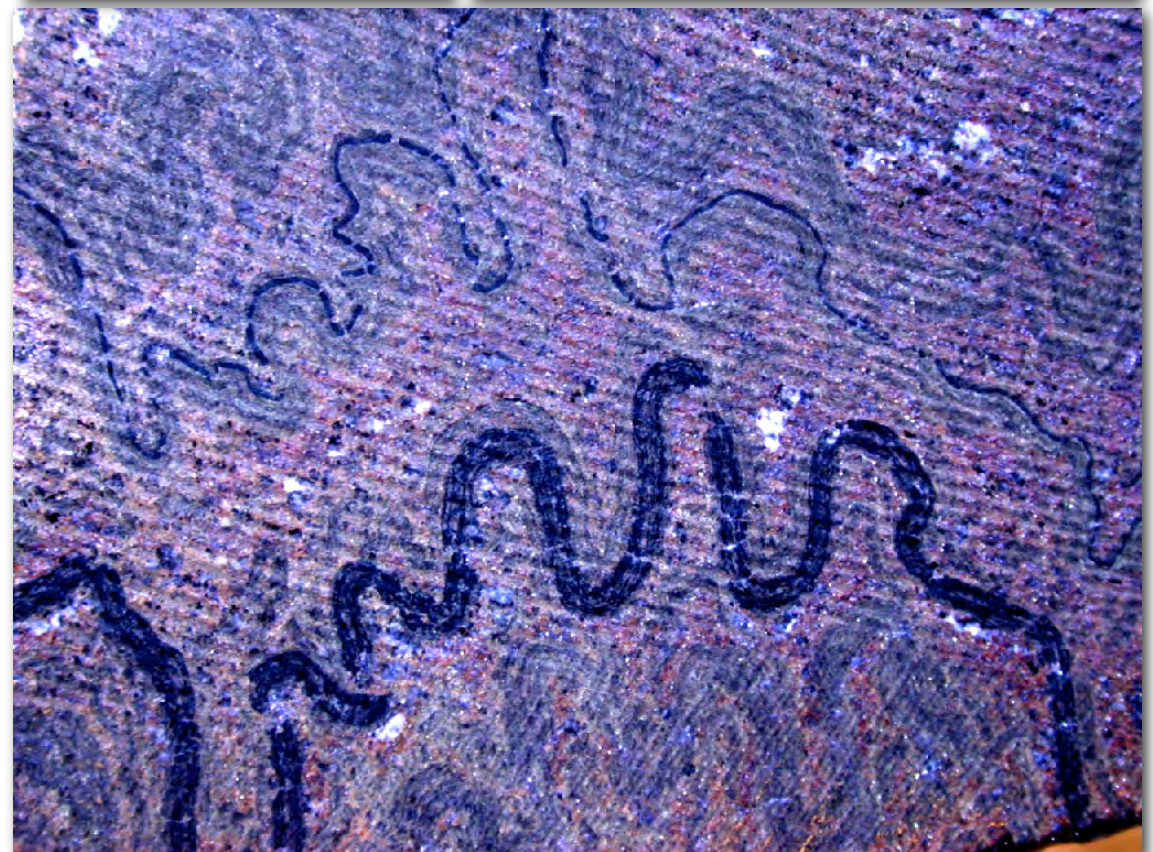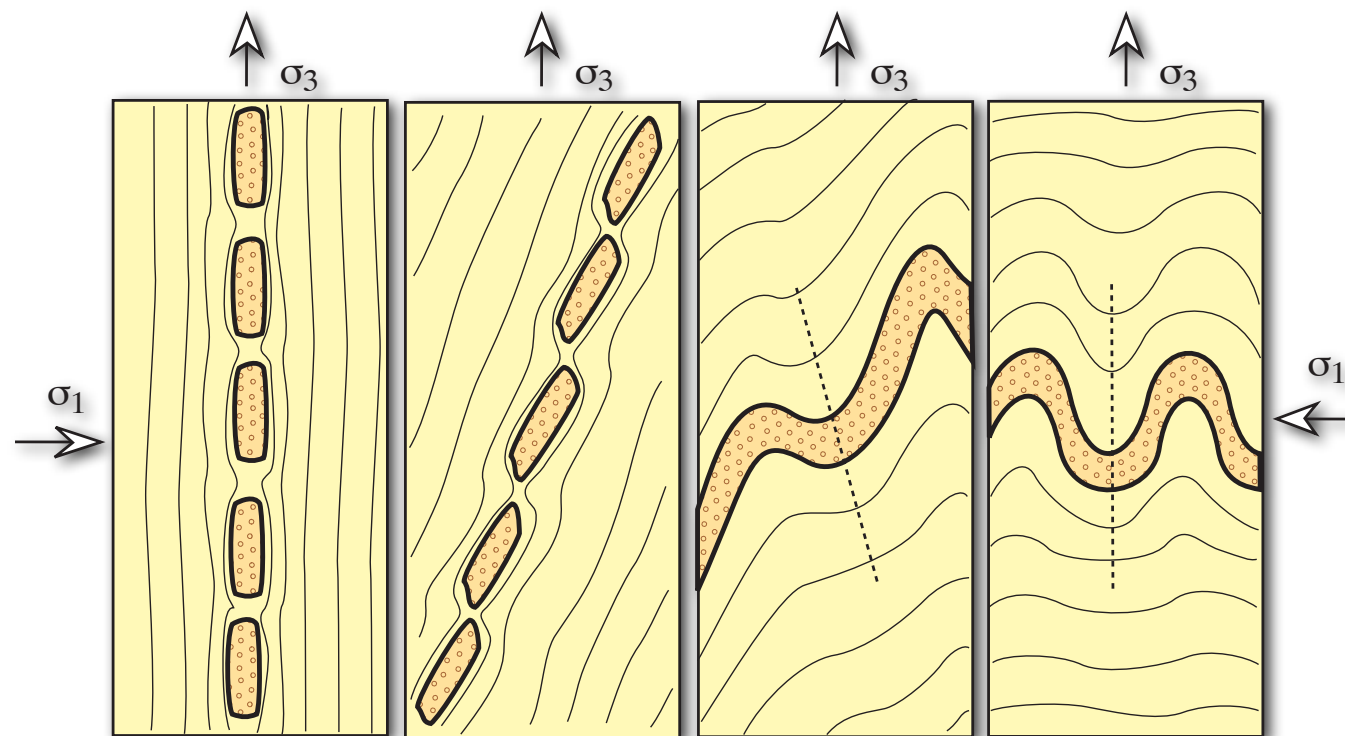
Photos on top right: Folds and boudins (Montagne Noire, France, Photo credit: P. Rey).
Photo at bottom right: Folds and boudinage in ore body, the picture is ca. 8 cm long (Mt Isa, Queensland, Photo credit: P. Rey)

CHAPTER 6
Ductile Shear Zones

The Ampahiny ductile shear zone in south Madagascar (S24º30′ - E44º56′, north is to the right, length of the picture is ~50 km) is a trans-lithospheric shear zone which has accommodated the amalgamation of Gondwana through collisional processes 650 to 600 Ma ago. The rock body in the middle is a Neoproterozoic anorthosite. The shear fabric wraps around this anorthosite developing a sigmoid pattern, which suggests a top to the north sense of shear. However the stretching lineation is sub-vertical, which suggests a complex polyphased history of deformation.

Definition

Ductile shear zones develop as a result of slow, progressive deformation over long periods of time. Deformation is typically continuous and develops without macro-scale fracturing. Rocks stretch but do not break. Deformation is said to be ductile as opposed to brittle. Shear zones are therefore fundamentally different to brittle faults, which involve sudden mechanical instabilities that explain earthquakes.

While faulting and fracturing typically develop at relatively low temperatures (<250°C), and/or high-strain rates, and/or high-deviatoric stresses, ductile shearing is the mode of deformation at higher temperatures (>250°C), low strain-rates, and low deviatoric stresses. Ductile shear zones tend to form bands of localised deformation.

Ductile shear zones are scale-independent and form from millimeter-scale shear bands, to transcontinental shear zones many tens of kilometers in thickness.

Gallery 6.1 Ductile shear zones

South part of Madagascar (North is to the right). White arrow points to the anorthosite massif shown on the first page of this chapter.

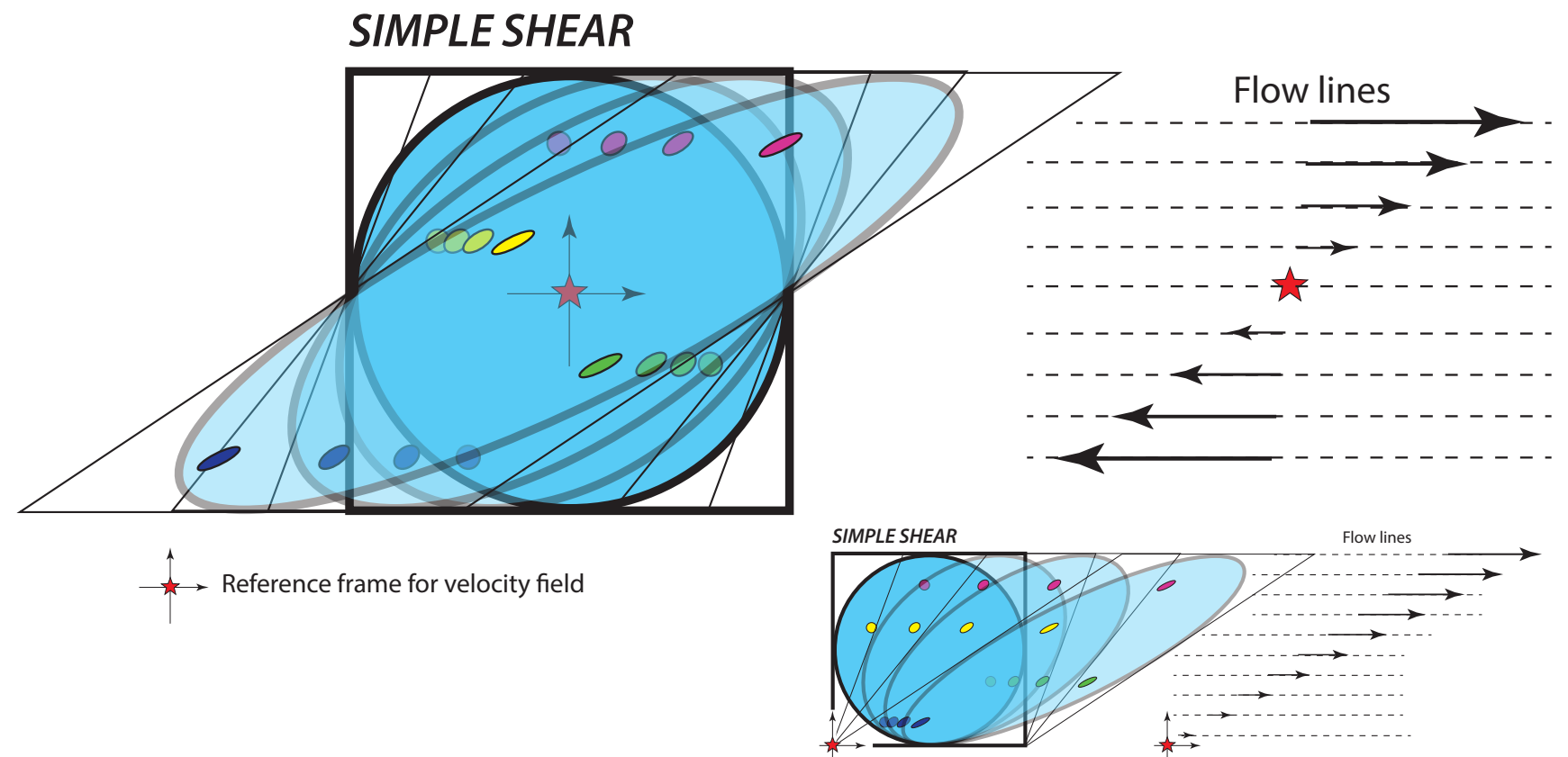# Ductile Shear Zones: Pure Vs Simple Shear Strain Regimes

Ductile shear zones develop as a result of shearing (*simple shear strain regime* as shown on the sketch at bottom right) or "squeezing" (*pure shear strain regime*) as illustrated on the sketch at the top right. In both cases, ductile shear zones form a localised band in which strain progressively increases toward the center of the shear zone, as shown on the conceptual sketches in which deformation is represented by passive markers. As a result, a foliation (i.e. a flattening plane analogous to the cleavage plane in folds), develops. Foliation and schistosity are often referred to as "planar fabric" and are labelled as: S_1, S_2, ... S_n, n numbering the phase of deformation.

Across simple shear ductile shear zones, foliation trajectories are asymmetric, as foliation forms at an angle to the main shear plane. As shear strain accumulates, this angle decreases as the foliation becomes progressively parallel to the main shear plane. The angle between the foliation and the main shear plane is a kinematic criteria as it indicates the relative sense of shear.

In a pure shear ductile shear zone, the foliation trajectories are parallel to each other across the shear zone, and remain parallel as strain accumulates through times.

Shear zone

Fault

S_1 trajectories

Shear zone
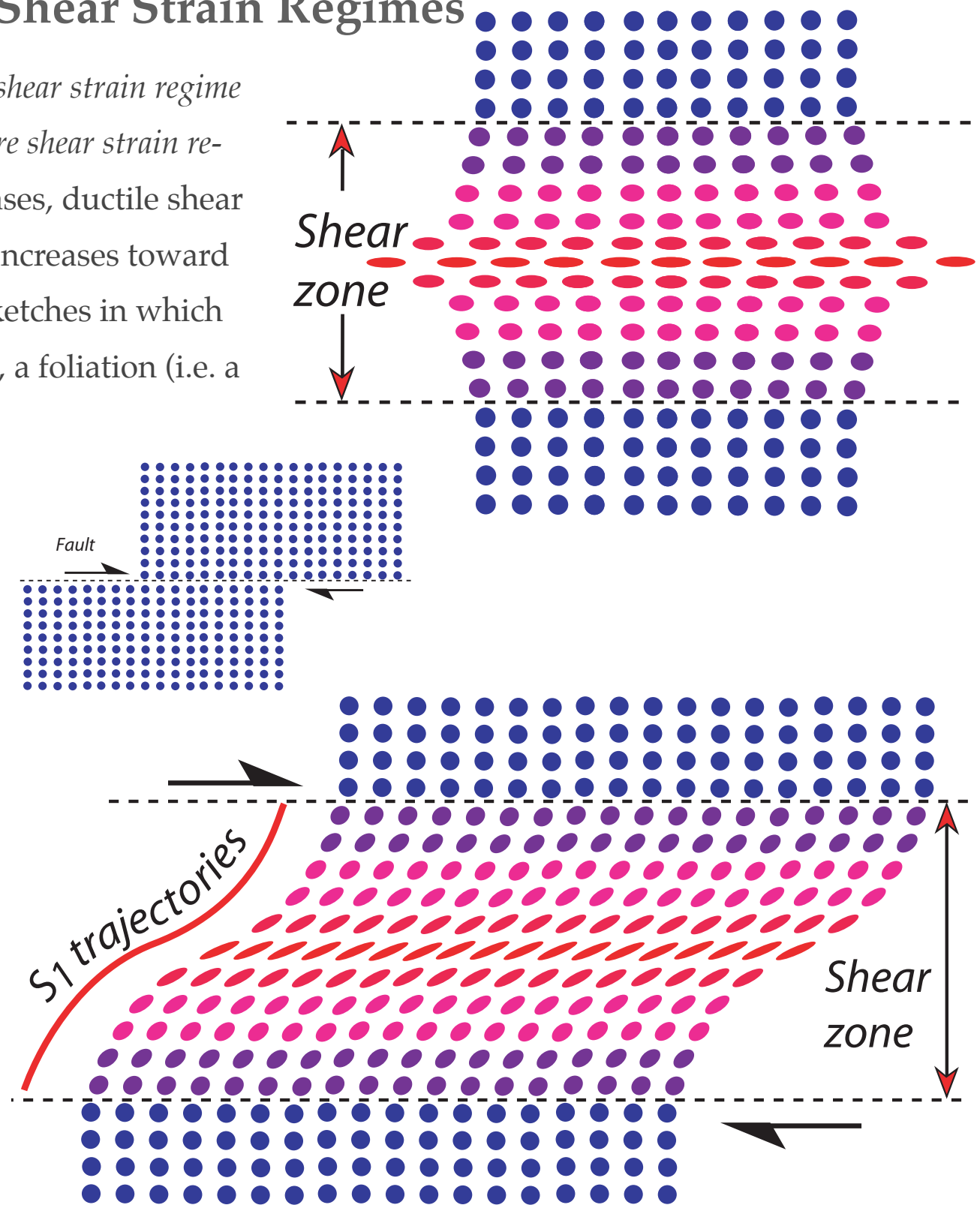
Flow lines during deformation: Flow lines represent the trajectories followed by physical points (e.g. molecules or atoms in the rock mass) during deformation. In the case of simple shear the flow lines are straight and parallel to each other. In the case of pure shear they are curved and they define a symmetric pattern. On both sketch the ellipses have the same shape, only different orientation.

NB: The flow field depends on the position of the reference frame.

SIMPLE SHEAR

Flow lines

Reference frame for velocity field

SIMPLE SHEAR

Flow lines
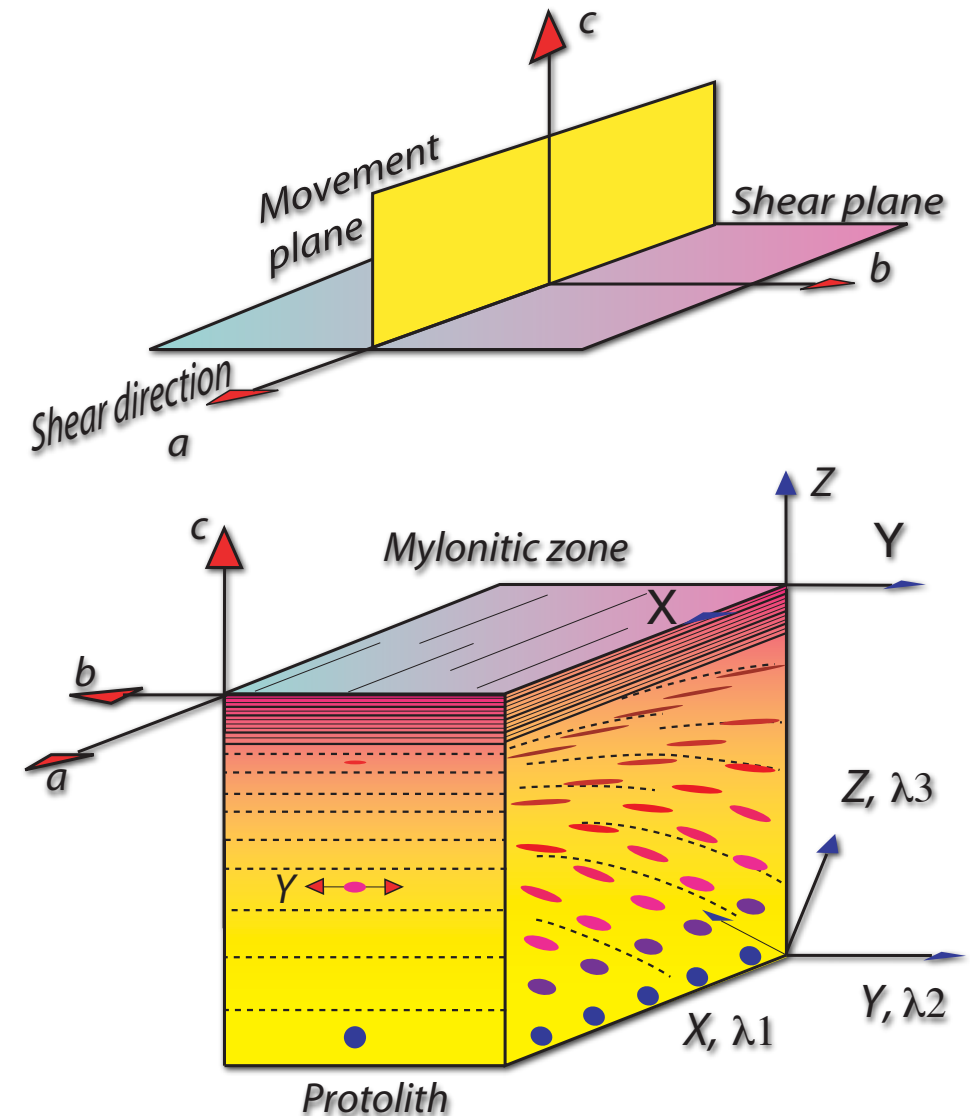
PURE SHEAR

Flow lines
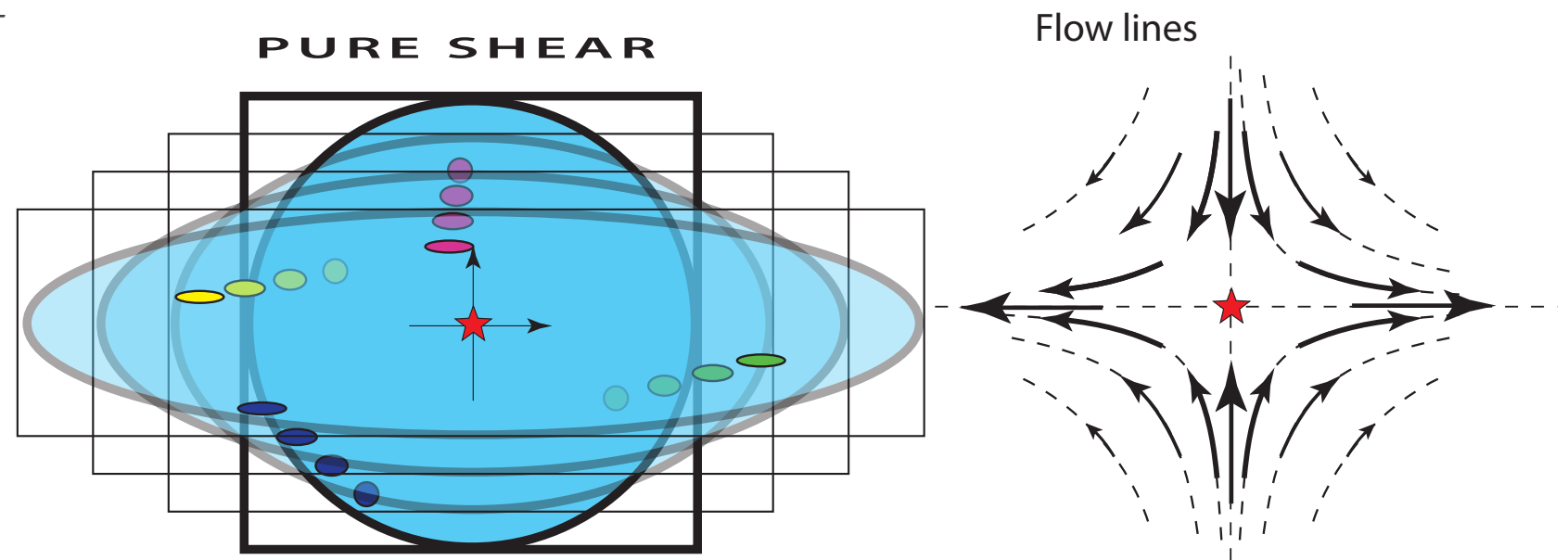
Ductile Shear Zone Descriptors

In shear zones, deformation induces a progressive reduction of the size of individual grains and minerals, as they recrystallize into many smaller grains. When most grains (70 to 80%) are no longer identifiable to the naked eye the shear zone is referred to as a *mylonite* or even *ultramylonite* for > 90% recrystallisation. In mylonites, clasts float into a fine-grained matrix made of recrystallized grains. Mylonite develops first along the central plane of the shear zone where deformation is the most intense. As shearing proceeds, the thickness of the mylonitic band and that of the shear zone increases. The *protolith* corresponds to the undeformed rock.

Two reference frames are used to describe shear zones:
The first relates to the finite strain framework. It consists of three orthogonal axes: X, Y, Z (or λ_1, λ_2, λ_3). Together, X and Y define the flattening plane (i.e. foliation plane). X is parallel to the direction of maximum elongation (i.e. the maximum stretching direction), therefore it is parallel to mineral and stretching lineations formed by the progressive rotation and stretching of minerals and mineral aggregates. Z is the pole of the foliation plane.

The second framework is related to the displacement field, it is called the kinematic frame. It consists of three orthogonal axes: *a*, *b*, and *c*. Together *a* and *b* define the shear plane in which *a* corresponds to the shear direction. This plane is parallel to the mylonitic core of the shear zone. The axes *a* and *c* define the movement plane. This plane is perpendicular to the shear plane and parallel to the shear direction. The kinematic reference frame remains constant in orientation across the shear zone. In contrast, the finite strain reference frame progressively rotates around the axis Y to become parallel to the kinematic frame in the mylonitic core of the shear zone. The sense of rotation is a kinematic criteria.

Shear Criteria & Kinematic Analysis

Pressure shadows

Kinematic analysis is the art of figuring out the relative sense of shear in shear zones, when they involve a strong component of simple shear.

One class of kinematic criteria is based on the asymmetric flow and asymmetric finite strain that result from the presence of minerals and mineral clasts more resistant to deformation than the enclosing matrix. Upon shearing, zones of *pressure shadow* and zones of *pressure concentration* develop around these clasts. As fluids and their solutes migrate from regions of high to low pressure, mineral growth occurs in pressure shadows, while dissolution affects zones of pressure concentration. The asymmetry of *crystallization tails* around clasts depends on the sense of shear, and as such can be used as kinematic criteria. Crystallization tails develop around a clast being progressively dissolved. The rate of dissolution is faster for a sigma pressure shadows than for a delta pressure shadows.

Pyrite grains in shale are an extreme case of hard clasts embedded into a soft matrix. As pyrite rotates during shearing, space incrementally opens up at the pyrite's margins. This space is filled with fluids from which minerals such as quartz and calcite precipitate often in a fibrous form.

Gallery 6.1 Pressure shadows and crystallization tails

Caledonian (ca. 400 Ma) eclogite facies ductile shear zone (Holsnoy, Norway, photo credit P. Rey)

1 of 12
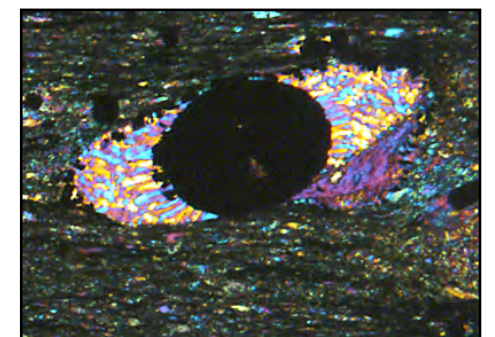
@ Jörn Wichert

Helicitic inclusions

Helicitic inclusions in porphyroblasts- Some minerals like garnet and staurolite can grow during shearing. These so called *porpyroblasts,* as opposed to *porphyroclasts* which are old fragments of mineral, act as a tape recorder for deformation. As they grow, porphyroblasts "freeze" the schistosity that develops in the surrounding matrix. An internal schistosity is recorded, the pattern of which reflects the kinematic of the shearing.


Rotation of syn- kinematic minerals
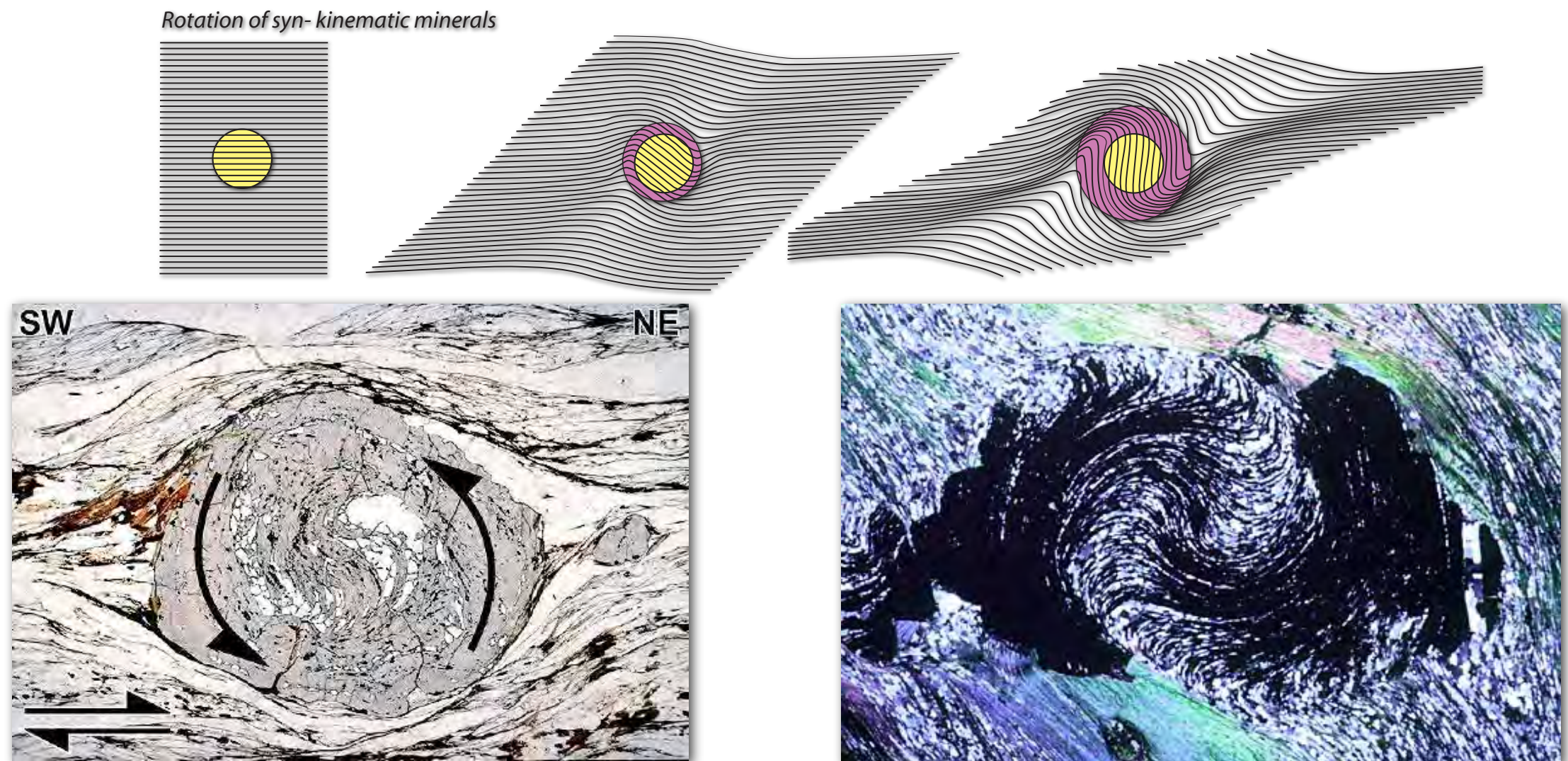
@ Pierre Dèvez: Syntectonic garnet in a metapelite. Helicitic inclusion trails indicate top-to-the-SW rotation. Shear sense criteria in the matrix indicate an opposite shearing associated to a more recent shearing event.

@ Scott Johnson: Syntectonic garnet porphyroblast in a metapelitic rock. Helicitic inclusion trails indicate clockwise rotation of the garnet during shearing.

Mica fish

Mica fish- Mica, usually muscovite, progressively rotate during deformation. At some stage their cleavages, which correspond to a plane of easy slip, become sub-perpendicular to the direction of maximum stress. When this happens, the cleavage plane of the mica becomes locked. Deformation jumps to the margins of the mica where micro-shear bands initiate. These micro-shear bands progressively lengthen to become shear bands.
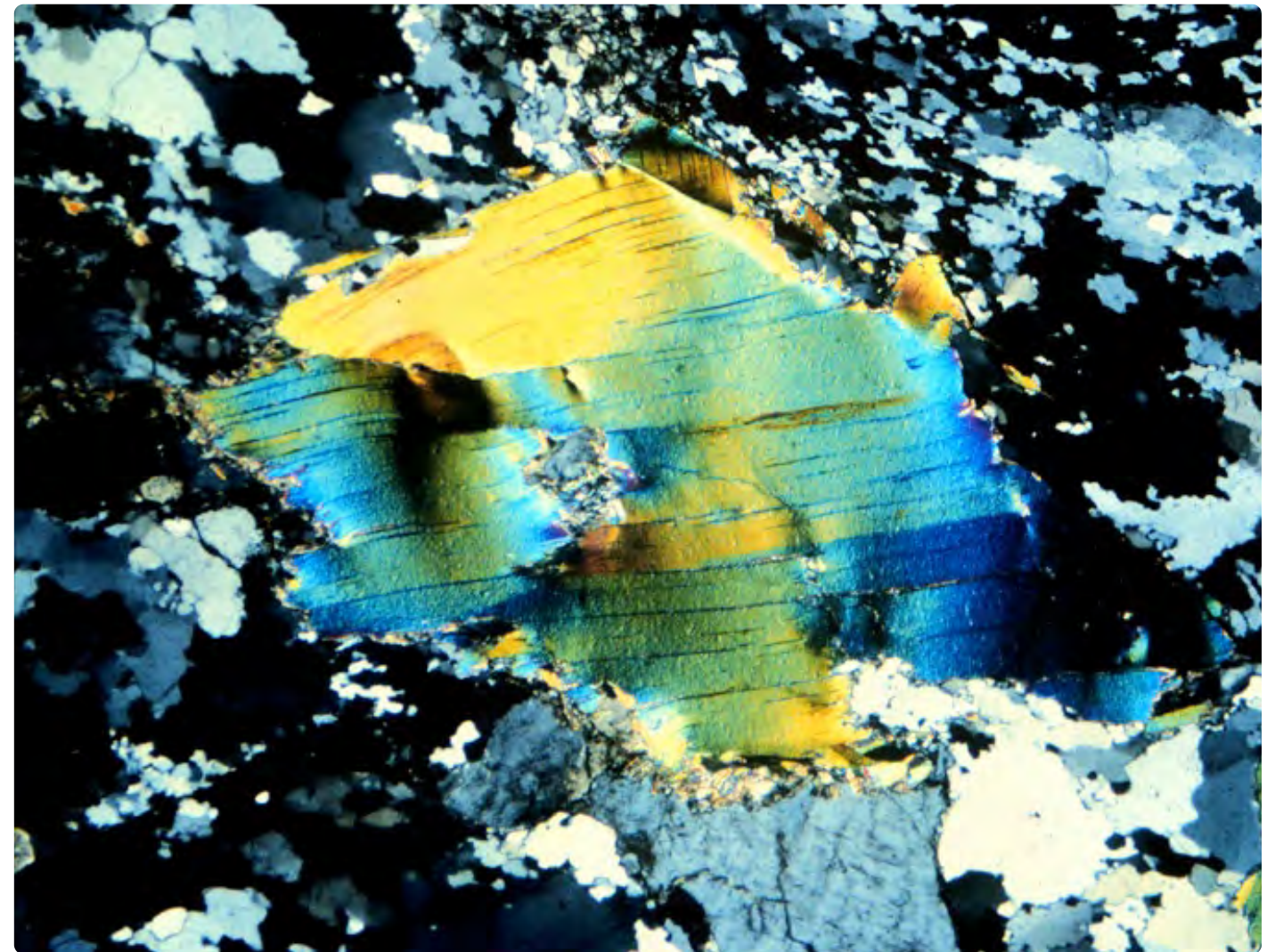
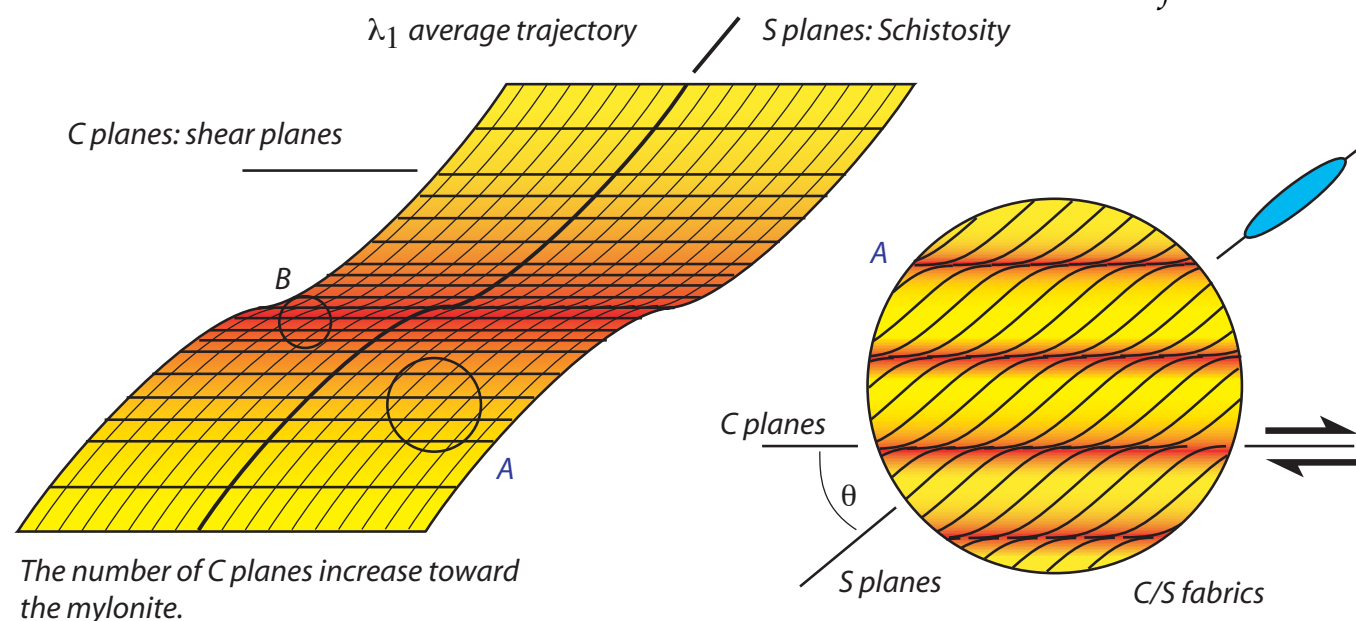*Large muscovite grain in a leucogranite from the French Massif Central (Photo credit: P. Rey)*

Mica fish
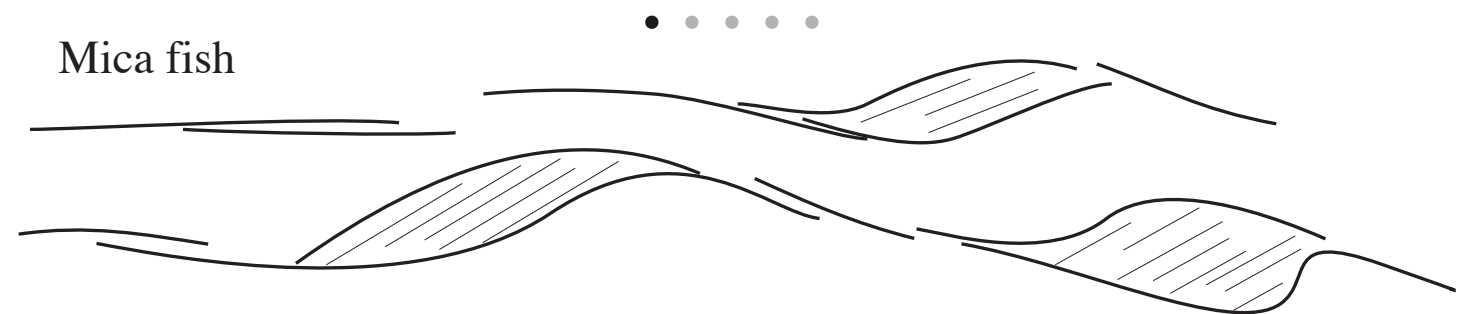
C-S and C-S-C' shear fabrics

C-S and C-S-C' shear fabrics - A class of kinematic criteria is based on the obliquity between the kinematic and the finite strain reference frames. Across a shear zone, the foliation - or S plane - (S for schistosity) rotates to parallel the shear plane - or C plane - (C for the french word *cisaillement*). As shearing increases the angle between S and C decreases (see below). This rotation is consistent with the kinematic of the shear zone i.e. a clockwise rotation of the S plane implies a clockwise sense of shear on the C plane.

In the mylonitic core, where C and S are sub-parallel, a new generation of shear plane called C' may develop with a small angle (10-15°) to the C plane. This is illustrated in the sketch showing a C/S/C' mylonite.
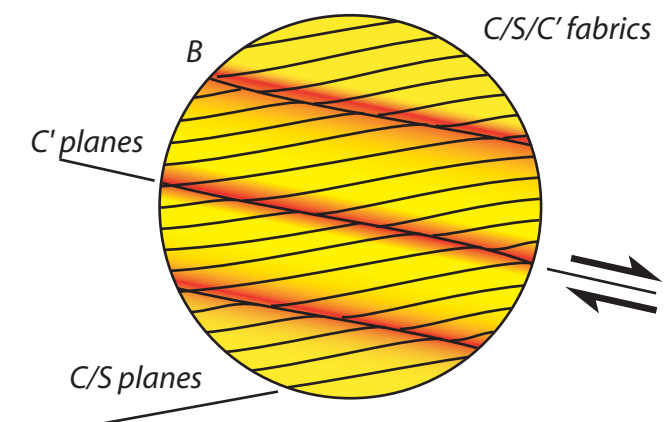
Gamma ~1, foliation clearly visible (leucogranite du Gueret, North part of the French Massif Central, Photo credit: P. Rey).

λ₁ *average trajectory* — *S planes: Schistosity*

C planes: shear planes

B

A

The number of C planes increase toward the mylonite.

A

C planes

θ

S planes

C/S fabrics

C/S/C' fabrics

B

C' planes

C/S planes

A Little Quiz...

1/ What is a shear zone?

2/ Define the concept of pure shear strain regime?

3/ How can one tell when a shear zone is the result of a simple shear strain regime?

4/ Define the following terms: Mylonite, Protolith.

5/ What is the difference between the movement plane and the shear plane?

6/ Define the strain reference frame and the kinematic reference frame.

7/ What are kinematic criteria?

8/ Page 66 shows a ductile shear zone, what is its kinematic?

9/ What is an SC mylonite?

10/ Why does SC mylonite cannot form during a pure shear strain regime?

Review 6.1

Question 1 of 3

Ductile shear zones develop at:

✓ **A.** Temperature $> 250°C$

○ **B.** Low strain rates

○ **C.** High differential stress

○ **D.** Only during simple shear

◀   Check Answer   ▶

Foliations and Lineations

During deformation, rock forming grains and minerals change their orientation and shape, giving deformed rocks organized planar and linear structures called fabrics. Foliation (the result of flattening) and lineation (most of the time the result of elongation) are planar and linear fabrics respectively that record the finite strain.

Definitions

Planar fabrics resulting from deformation are called schistosity or foliation. Foliation results from the ductile flattening of grain aggregates (e.g. quartz), and/or the change in orientation of tabular minerals (e.g. micas), and/or the anisotropic growth of newly formed minerals. Often, this flattening is not iso-tropic, i.e. a sphere is not flattened into a circular pancake but into an elliptic one. In this case, the foliation plane carries a linear fabric called a lineation. This lineation is due to the stretching of grain aggregates (stretching lineation), the re-orientation of elongated minerals such as amphibole, rutile (mineral lineation), and the preferential growth of newly formed minerals. Usually, but not always, stretching lineation corresponds to the direction of maximum stretching (X) within the foliation plane (XY). In the field, the orientation of a foliation plane is determined by its strike, its dip and its dip direction, and the orientation of lineation by its plunge and plunge direction.

Schistosity fabric in a finely bedded sedimentary rock (psammite, north of Broken Hill, NSW, photo credit P. Rey).

● ● ● ● ●

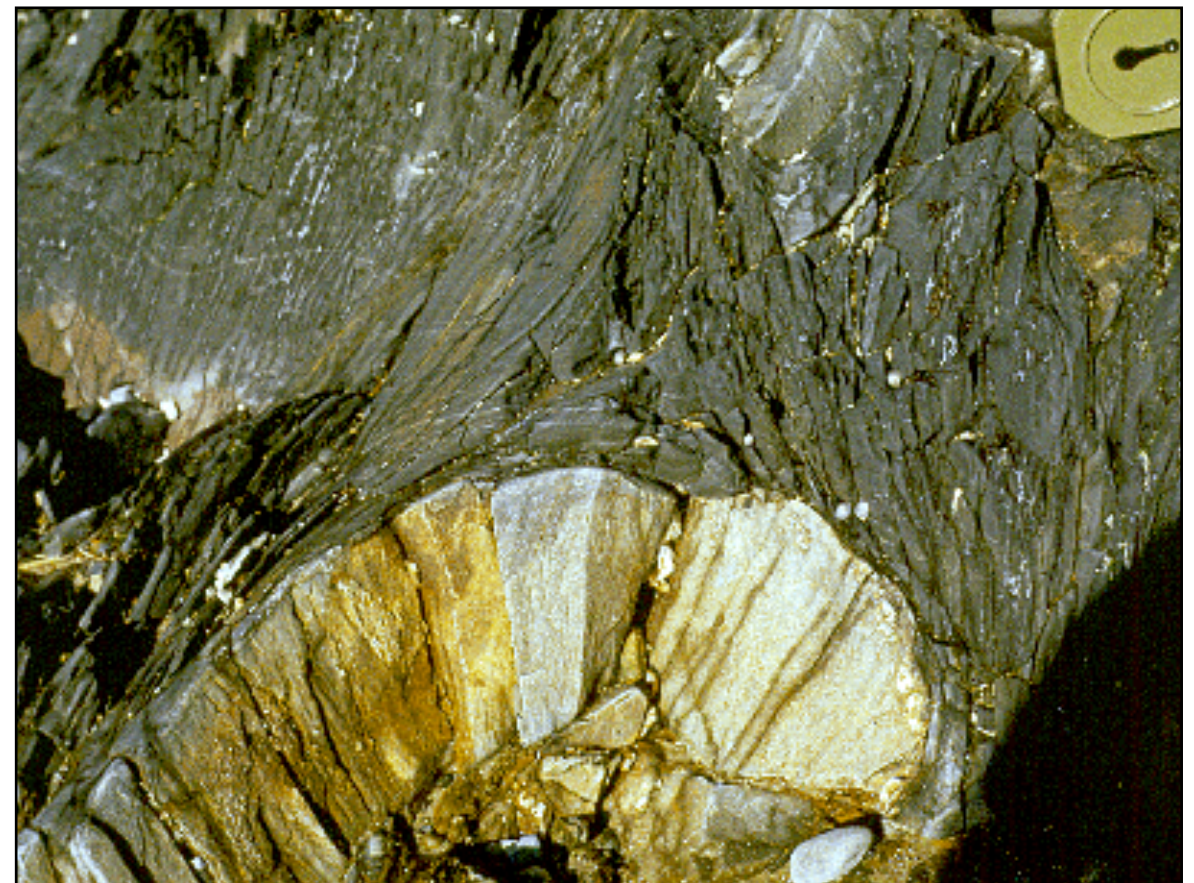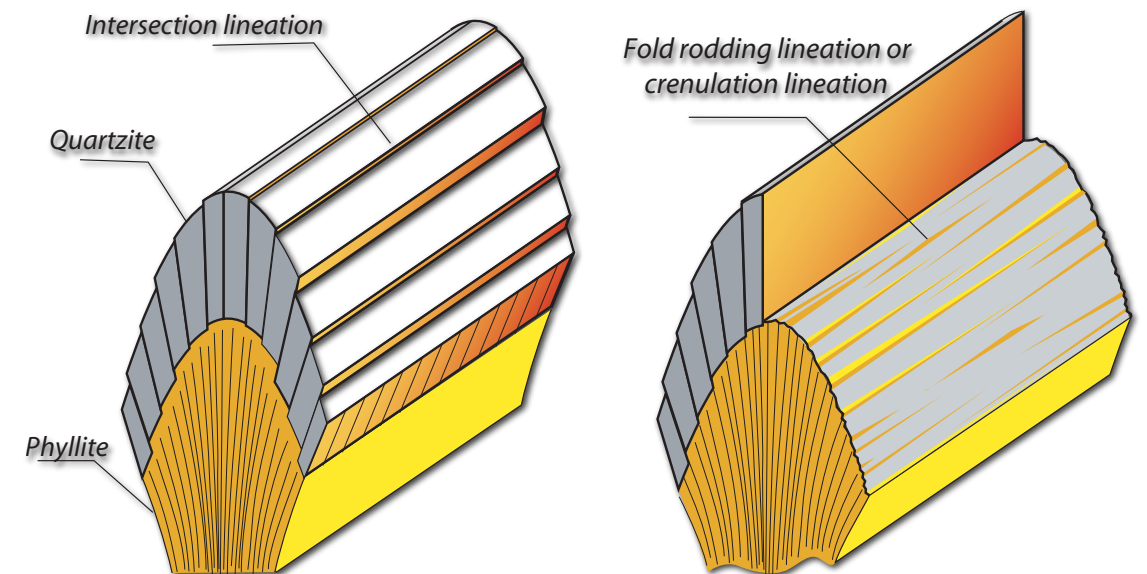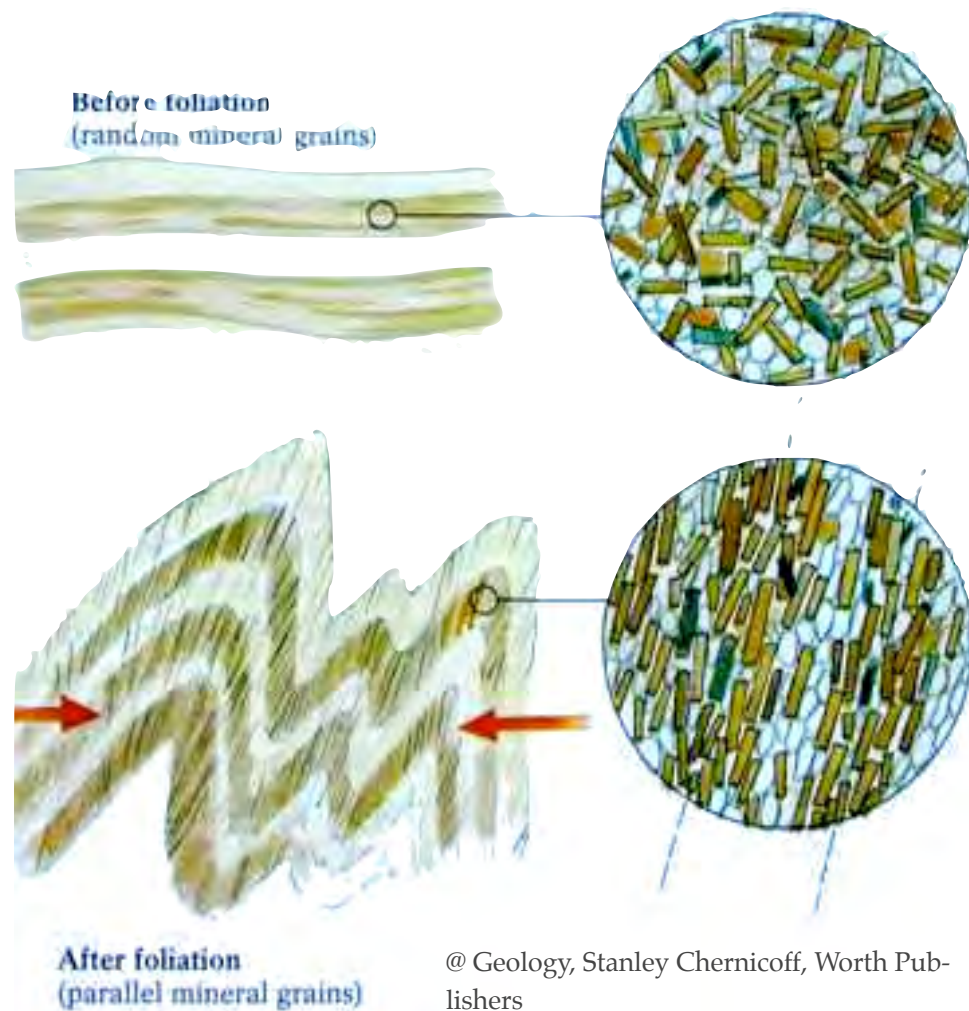The pictures in the gallery 7.2, illustrate an example of strong linear fabrics. The outcrop shows a deformed rhyolite (felsic volcanic rock equivalent to granite) of the Wyman formation (ca. 3.3 Ga) in the east Pilbara (WA). Such a strong linear fabric is referred to as a constrictional fabric. In the field, the attitude of a lineation is fully characterized by its plunge (angle from a horizontal surface down to the lineation) and plunge direction (azimuth of the imaginary vertical plane carrying the lineation).

The outcrop shows a deformed rhyolite (felsic volcanic rock equivalent to granite) of the Wyman formation (ca. 3.3 Ga) in the East Pilbara (WA). This outcrop shows a very strong steeply dipping linear fabric, and a rather weak to very weak flattening plane (Photo credit: P. Rey).

● ● ● ● ● ●

Foliation & Lineation Forming Processes

1/ *Folding* - During folding, grains and minerals are re-oriented and flattened into a planar fabric called axial planar fabric. This fabric may "refract" (i.e., change direction) across beds (picture on the lower right). Where this tectonic fabric intersects a pre-existing fabric, such as a bedding plane, intersection lineations form. These lineations run parallel to fold closures.


Intersection lineation

Quartzite

Phyllite

Fold rodding lineation or crenulation lineation

Before foliation (random mineral grains)

After foliation (parallel mineral grains)

@ Geology, Stanley Chernicoff, Worth Publishers

2/ *Shearing* - Shearing leads to the flattening and stretching of rocks forming grains and minerals. As temperature/depth increases, shearing results in more pervasive/continuous fabrics. At relatively low temperature (250-450°C) shearing is more localized and strong minerals such as feldspar are brittle. At relatively high temperature (>600°C) strain is more homogeneous as feldspar becomes ductile.
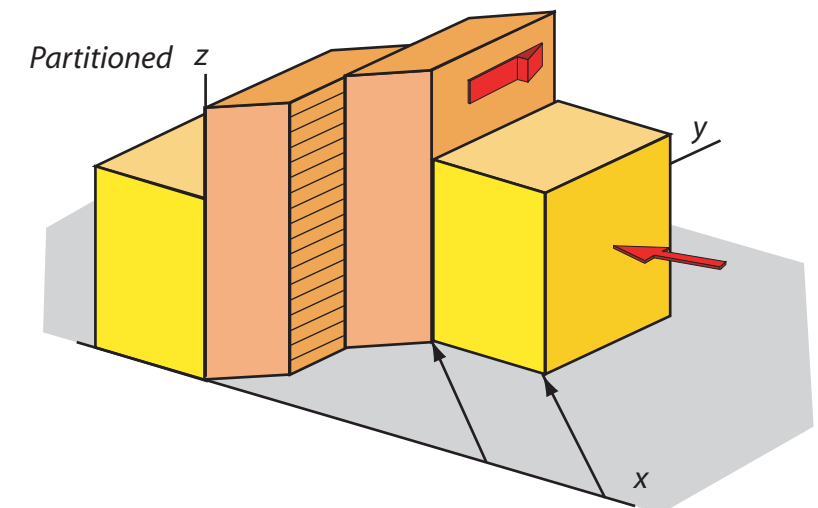
Upper Structural level

Discontinuous

Heterogeneous

Intermediate

Continuous

Lower

S

C

$\lambda_1\lambda_2$

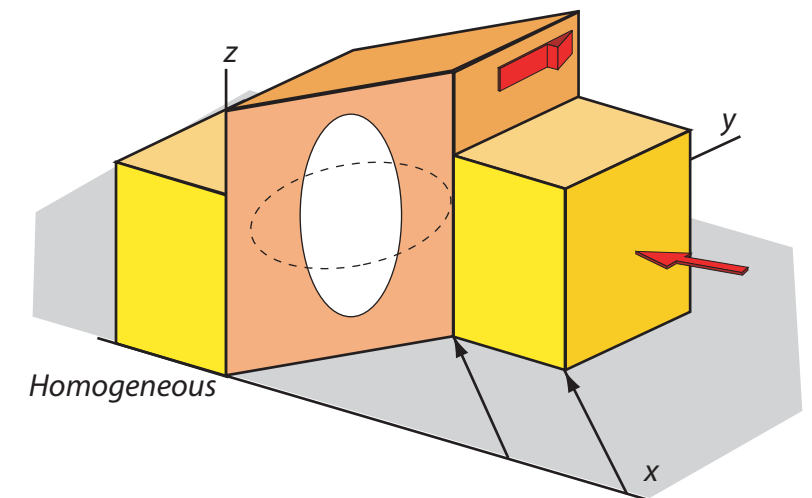$\lambda_1\lambda_2$

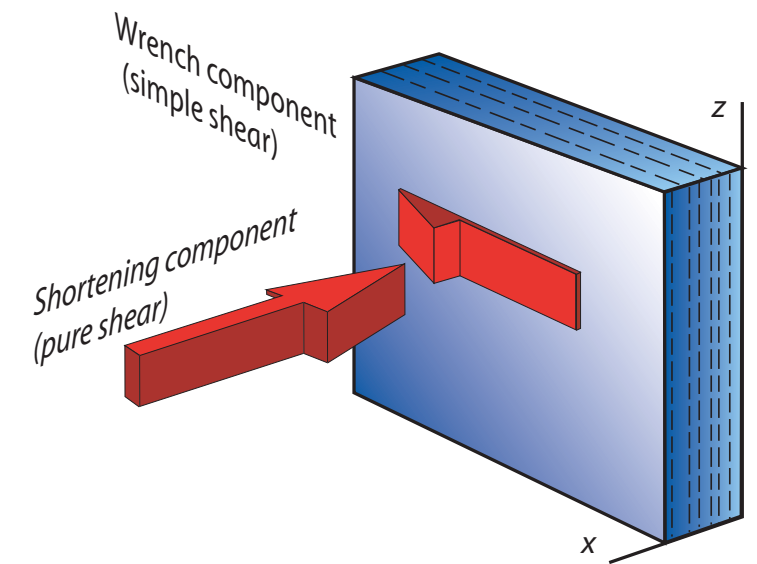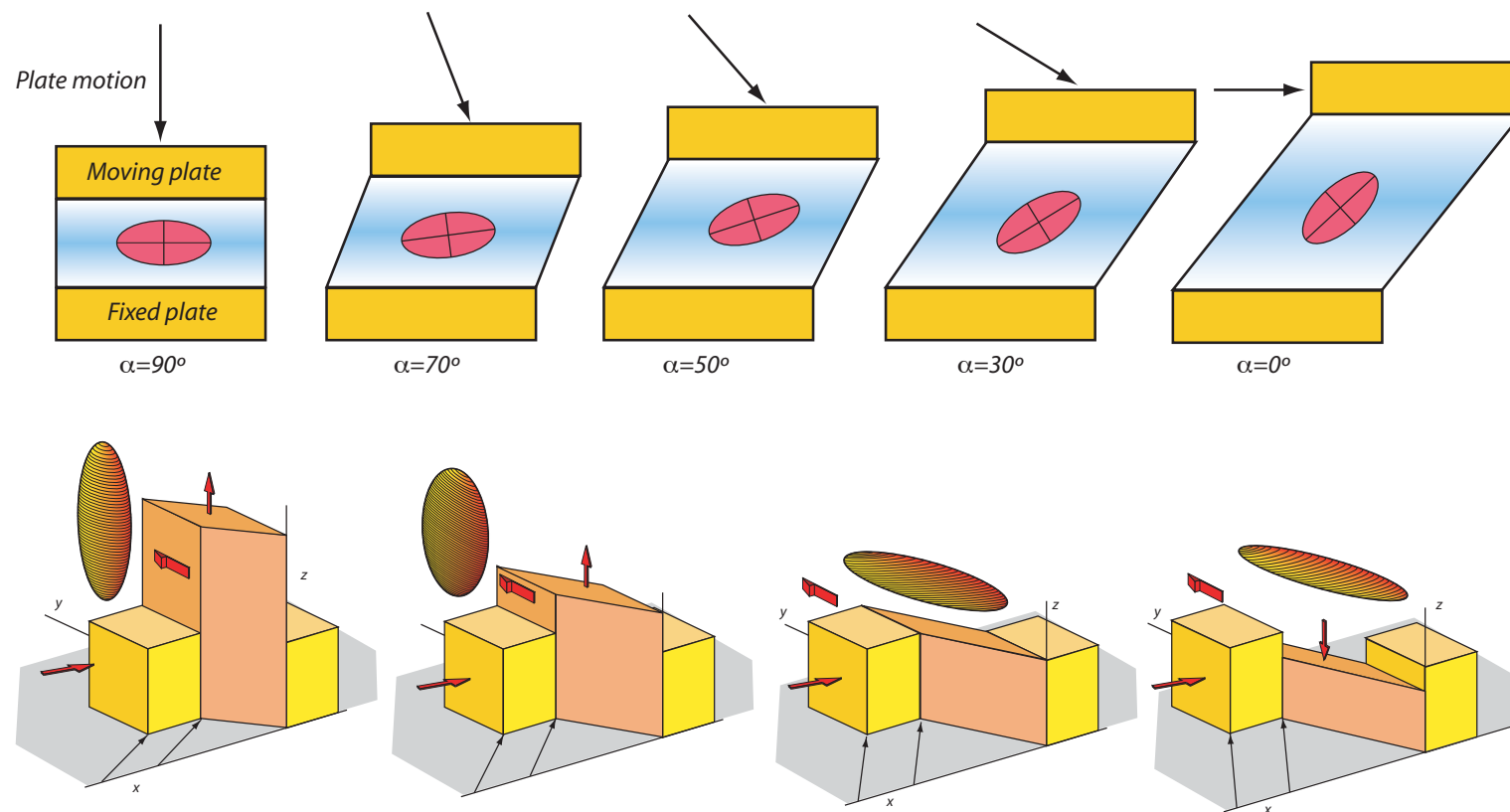**Gallery 7.3** Shear foliation and lineation

Shear zone with pressure shadows around garnets (Barberton Greenstone Belt, South Africa, photo credit: P. Rey).
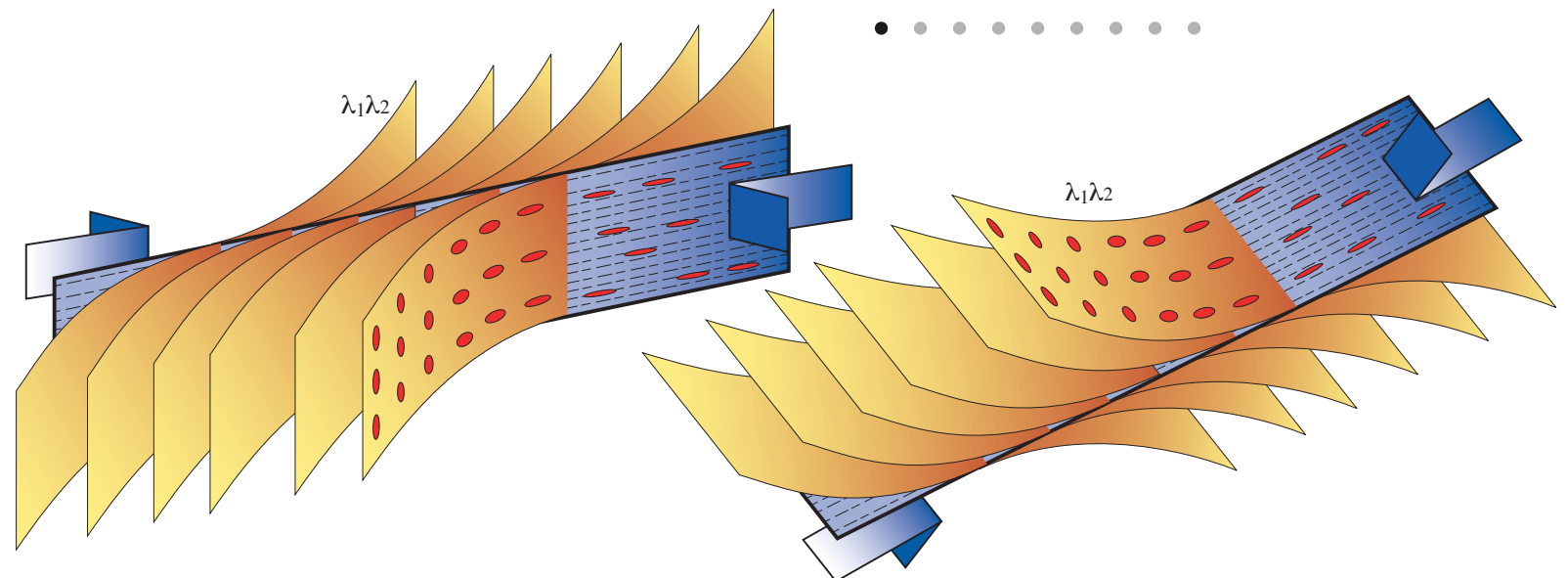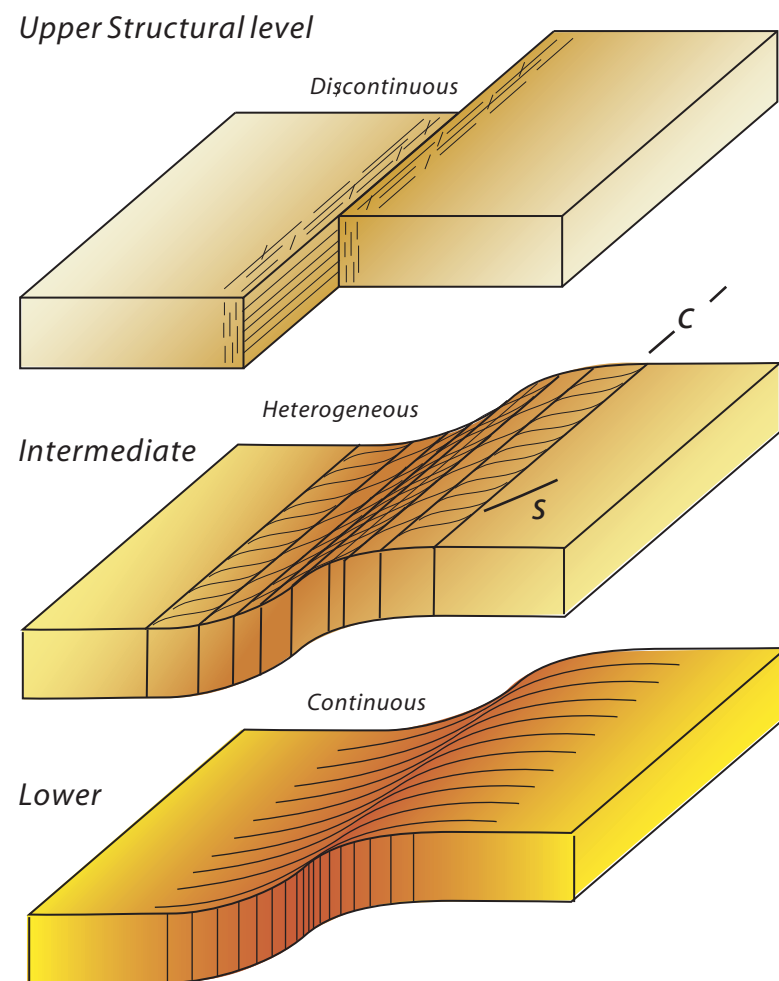
Strain partitioning and triaxial deformation

Simple shear and pure shear are only two end-members of an infinite gamut of shear in which both pure shear and simple shear are involved. In these cases, the relationship between the finite strain framework and the kinematic framework is complex and may lead to some unexpected results. For instance, in a triaxial shear, the stretching lineation can be strong to non-existent, and its plunge can be either vertical or horizontal depending of the relative contribution of each strain regimes. The process of strain partitioning can explain further complexity. Indeed, strain can be partitioned into regions experiencing contrasting strain regimes.

Wrench component (simple shear)

Shortening component (pure shear)

Plate motion

Moving plate

Fixed plate

α=90°   α=70°   α=50°   α=30°   α=0°

Homogeneous

Partitioned

3/ *Flow of magma* results in the rotation of magmatic minerals into magmatic foliations and lineations. Magmatic foliations do not say much about the direction of magmatic flow. Like any other tectonic fabric, magmatic fabrics only inform about finite strain (i.e. X,Y and Z axes of the finite strain framework). In the field, it is relatively easy to distinguish between magmatic foliations and solid-state foliation. Magmatic foliations are best seen with a bit of distance from the outcrop. Indeed, as one looks closely at a scale a few times the average grain size, the foliation is hardly visible. Solid-state foliations are visible at all scales.
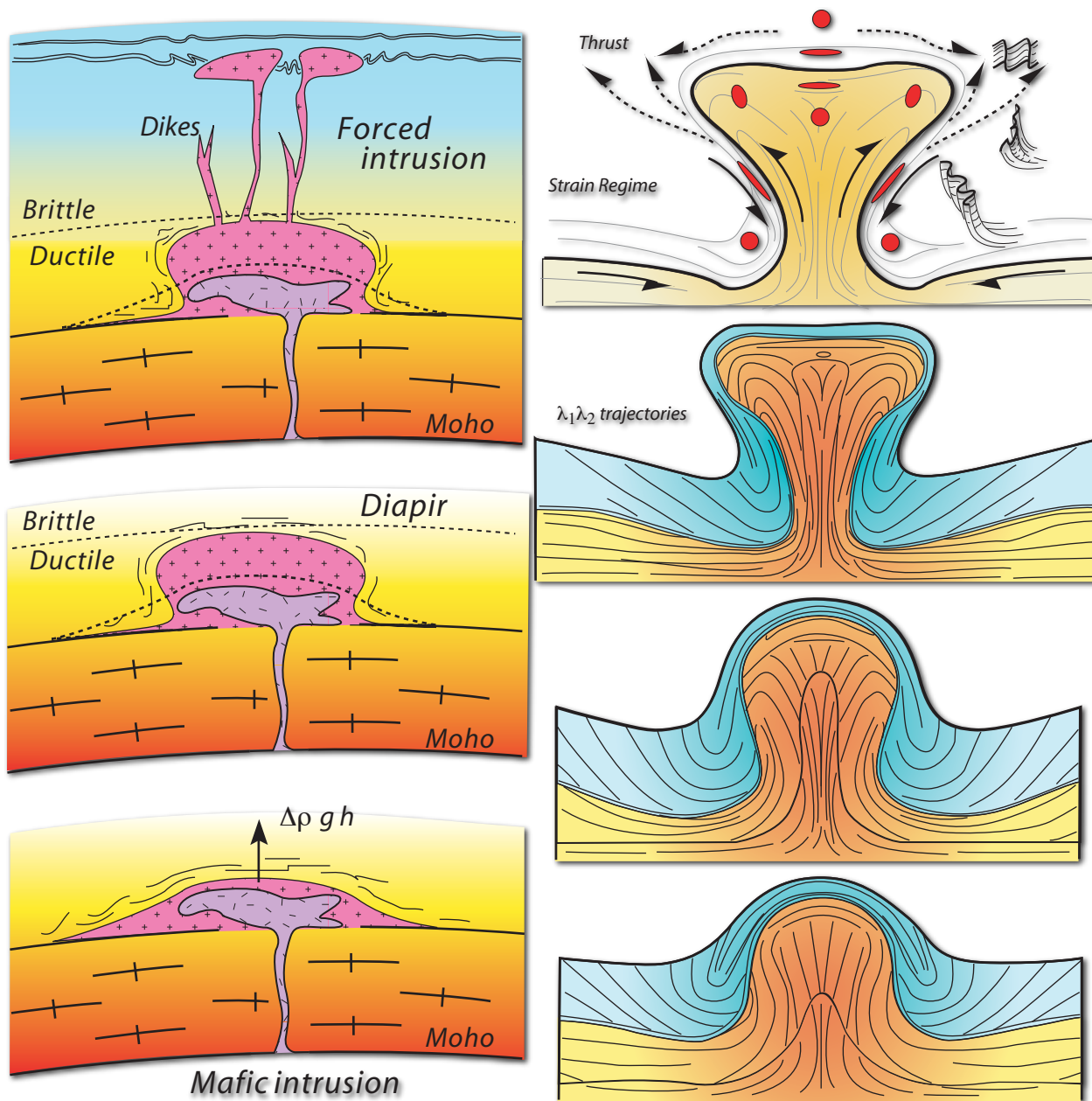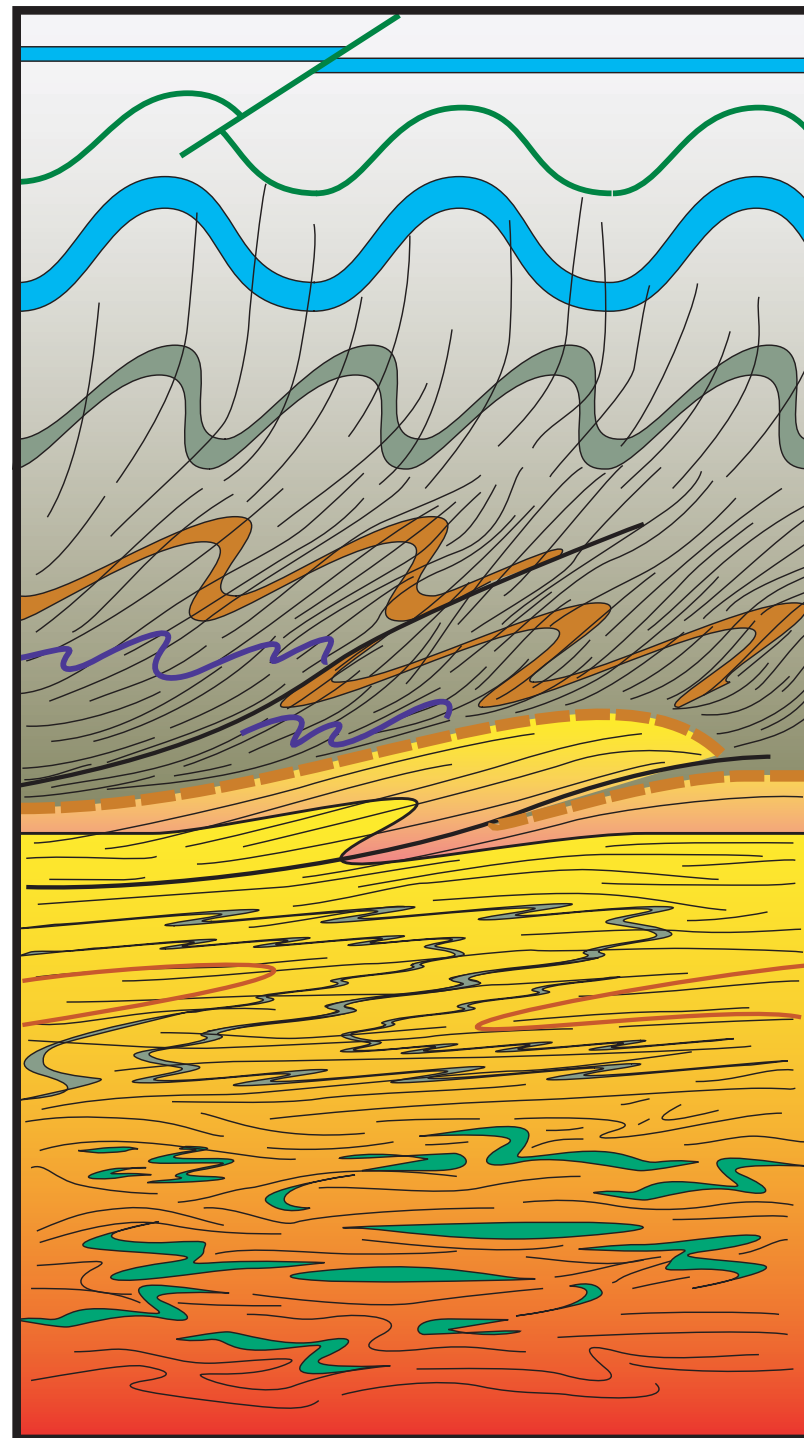

Gallery 7.4 Magmatic foliations and lineations
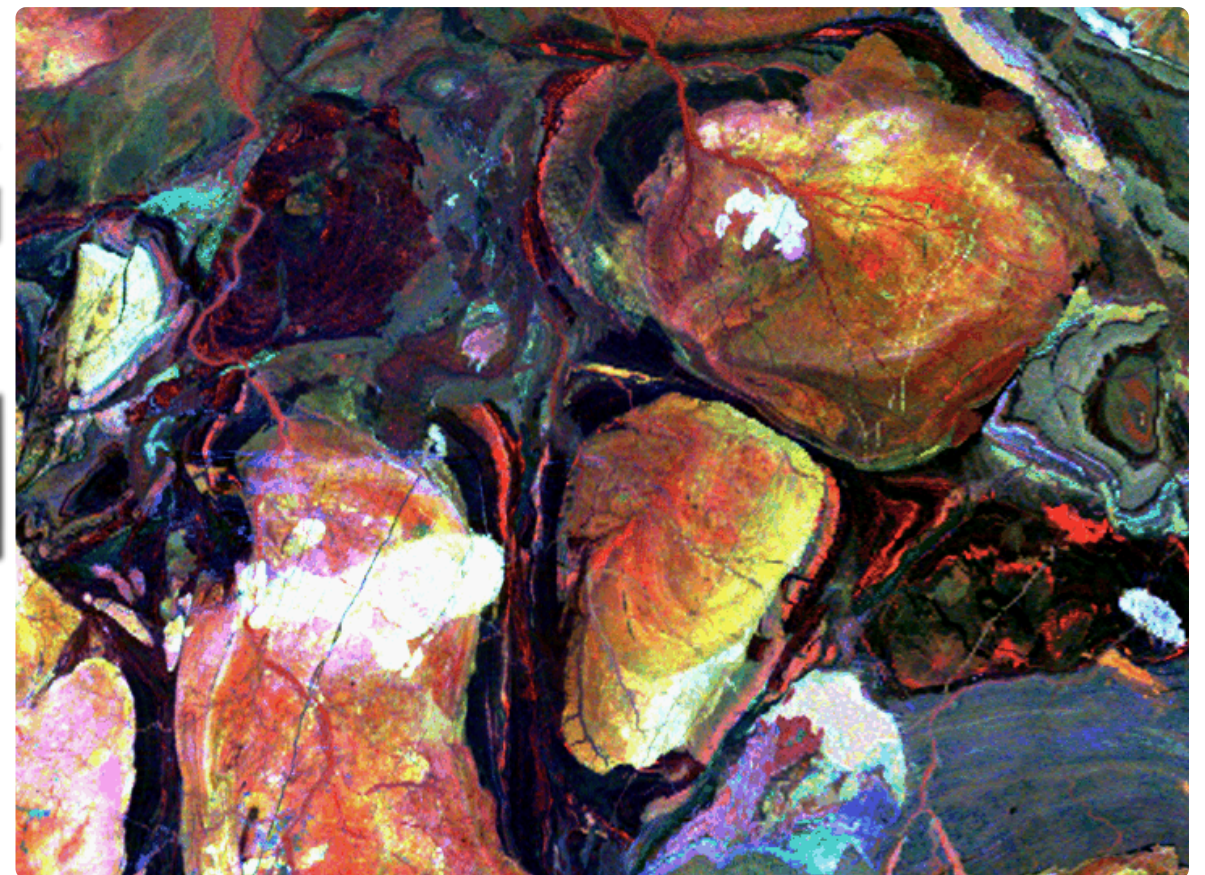
Radiometric image of the dome and basin pattern in the East Pilbara Craton (WA, Image: Geoscience Australia).

Fabrics In The Continental Crust

Faults

Upright, isopach folds.
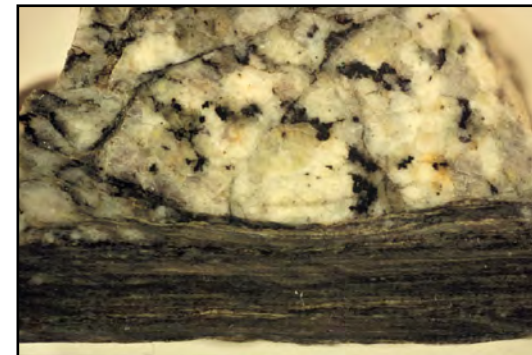
Schistosity front

Inclined, similar folds

Overturned, tight folds

Metamorphic nappes and recumbent folds

Tectonic transposition, recumbent isoclinal folds

Moho

As temperature increases with depth, rocks become weaker and their strength contrast, due to their constrating felsic-mafic composition, decreases. This explains the growing homogeneous character of tectonic fabrics with increasing depth.

Tectonite Classification

The relative strength (i.e. intensity) of the planar fabric and the linear fabric is used as a qualitative criteria to classify tectonites. S-tectonites are characterized by a strong and pervasive schistosity (or foliation) and no lineation. S-tectonites are the result of isotropic flattening in which there is one direction of shortening and multiple, radially symmetric stretching. In SL tectonites the foliation plane carries a mineral and/or stretching lineation. We distinguish S>L, S=L and L>S tectonics based of the relative intensity between the foliation and the lineation. Simple shear strain regime is often involved in the formation of SL tectonites. L-tectonites are characterized by a prominent lineation and no schistosity or foliation plane. L-tectonites form when there are multiple directions of shortening and one direction of stretching. Pure shear can produce S and L tectonites.
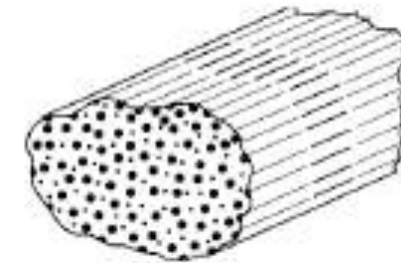
S - Tectonites

SL - Tectonites
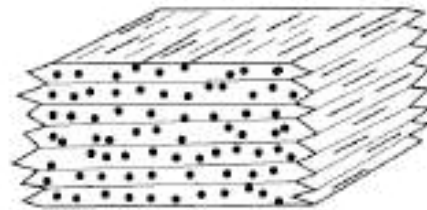
L - Tectonites

A Little Quiz

1/ In structural geology what is a "fabric"?

2/ What are the main fabric-forming processes?

3/ In a granite, how can we tell that a foliation formed at a temperature of about 400°C from one formed at a temperature of ca. 700°C.

4/ What is a tectonite?

5/ Define the three main types of tectonites.

6/ Can a ductile shear zone result in the formation of S-tectonites?

7/ What is the difference between an intersecting lineation and a stretching lineation?

L tectonite (Klondyke district, East Pilbara, WA, photo credit: P. Rey)

• ∘ ∘

8/ Are stretching lineations automatically linked to a foliation plane?

9/ Can stretching lineations pierce through a foliation?

10/ The attitude (ie orientation in space) of a foliation plane is measured through its strike-dip and dip direction. Draw a sketch showing the concepts of strike, dip and dip direction.

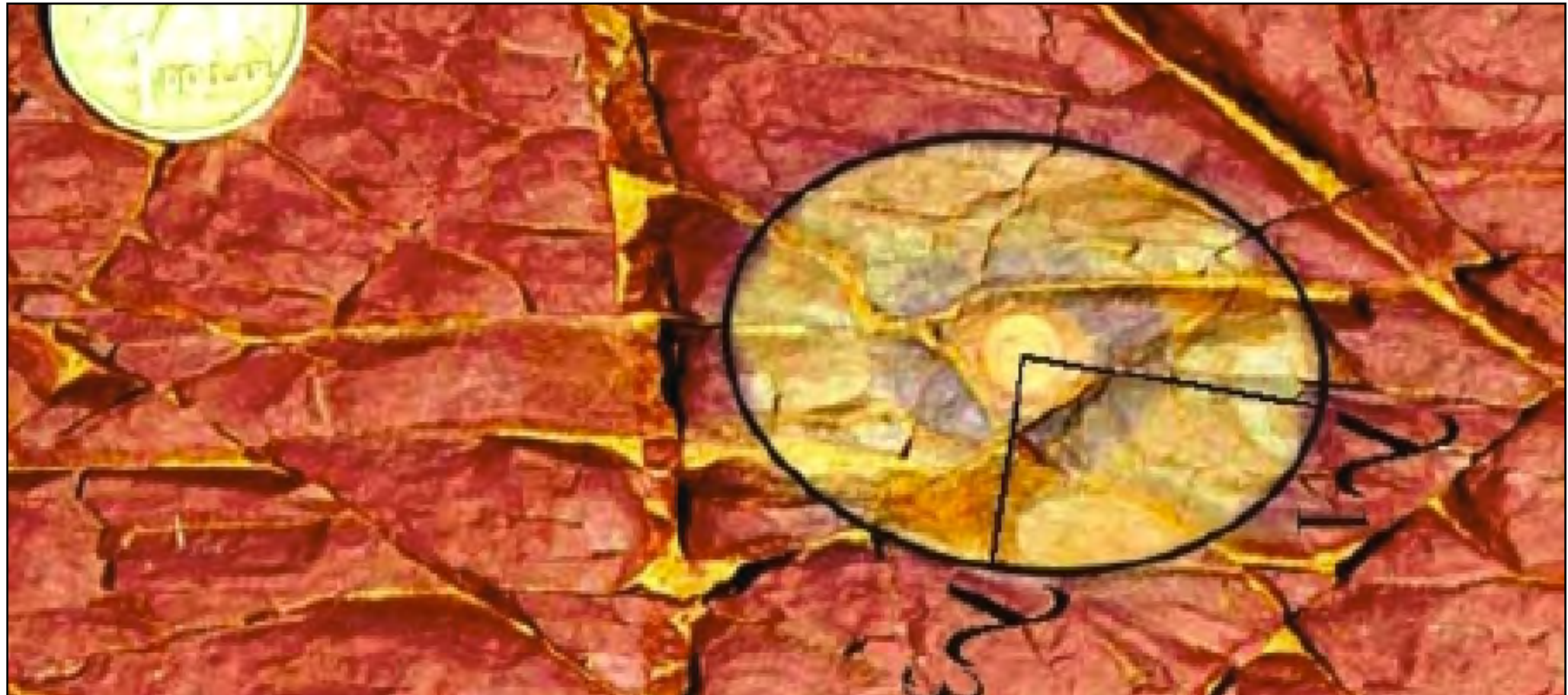# Strain and Strain Analysis

Structural geologists observe and document structures, such as faults and fractures, ductile shear zones, folds and boudins, foliations and lineations, not because they are pretty, but because they give information about finite and incremental strains. This chapter focuses on understanding progressive strain as a continuum process, and developing a strategy to quantify strain.

Continuous Vs Discontinuous Strain

Deformation is very often partitioned between discontinuities (faults and shear zones) that limit domains where deformation is continuous (cf. sketches below). Finite strain analysis aims at characterising the final state of deformation (the finite strain) in domains where deformation is continuous, whereas fault analysis characterises strain associated with discontinuities. Both analyses are necessary to fully characterise finite strain fields.

The analysis of faults and shear zones consists in documenting at the local scale their orientation (strike-dip-dip direction), thickness, geometry (single discontinuity plane, anastomosed network of shear zones or brittle faults, overlapping fault segments, etc.), their kinematic and kinematic history (relative sense of displacement through time), their amount of displacement, and finally to derive the orientation of the paleostress through time that resulted in the formation of the observed faults or shear zones.

Finite strain analysis consists of measuring, at the local scale, the direction of maximum shortening and lengthening, characterising the geometry of the strain (flattening, constriction, plane strain, etc.), determining the intensity of the strain, assessing the strain regime (pure shear or a simple shear), and when appropriate, determining the kinematic of the strain.

The sketch on the right shows a block before and after deformation. Shortening is accommodated by a reverse fault and rather complex but continuous internal deformation. To characterise the internal deformation of the two blocks we divide them into a number of small cells in which strain can be considered, at first approximation, as homogeneous (i.e. no strain gradient). When deformation is homogeneous, an imaginary sphere enclosed in each cell of the block before deformation is transformed into an ellipsoid from which strain can easily be characterised.

Finite strain analysis is only valid when strain is considered at a scale at which strain is homogeneous. The measurement of the characteristics of strain in a large number of cells across the entire block fully characterises the *Finite Strain Field*.
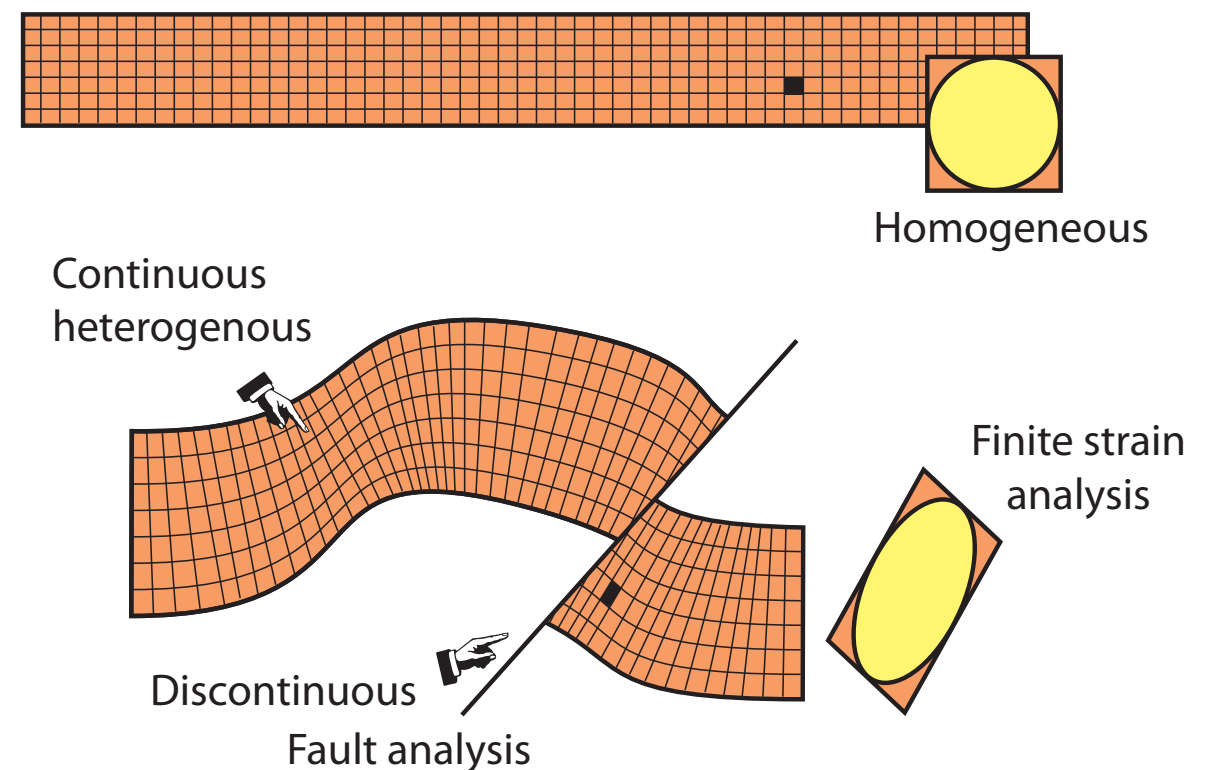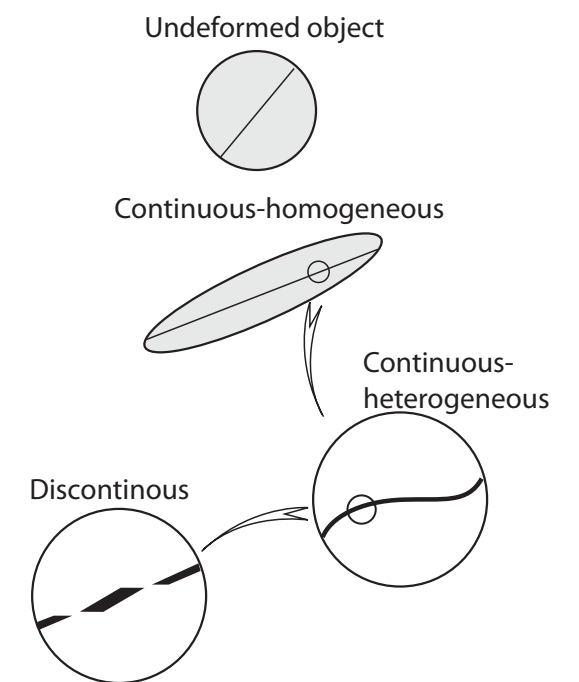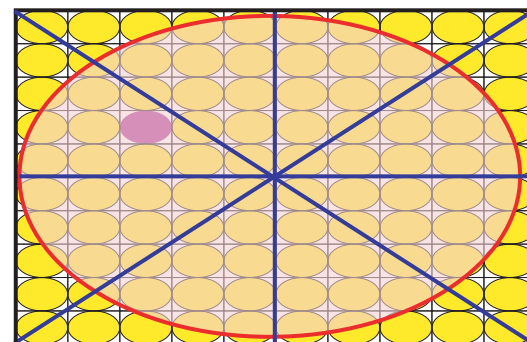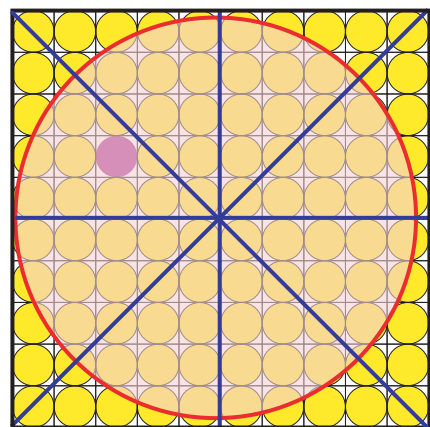
Homogeneous Vs Heterogeneous Strain

In structural geology, "*deformation*" and "*strain*" are not synonymous.The term "*deformation*" includes *translation*, *rotation*, *change in volume* and *change in shape*. The term "*strain*" refers only to the change in shape, a synonymous term would be "*distorsion*". Therefore, strain is only one component of deformation. In the field, translation, rotation and change in volume are difficult to assess, but not impossible. Strain analysis focuses on the characterization of the change in shape.

Geometric description of strain

Faults and fractures produce *discontinuous strain*. In this case, strain is assessed by the description of the network of faults and fractures as well as the rotation and translation of individual blocks. When strain involves "viscous" flow, strain is said to be *continuous* (cases B and C). If strain is the same at any location of the deformed body (i.e. no strain gradient as in B, it is said to be *homogeneous*. In contrast, if strain varies in shape or intensity from one location to another, then strain gradients exist and the strain is said to be *heterogeneous* (case C).

Initial state

B1 and B2Continuous homogeneous strain
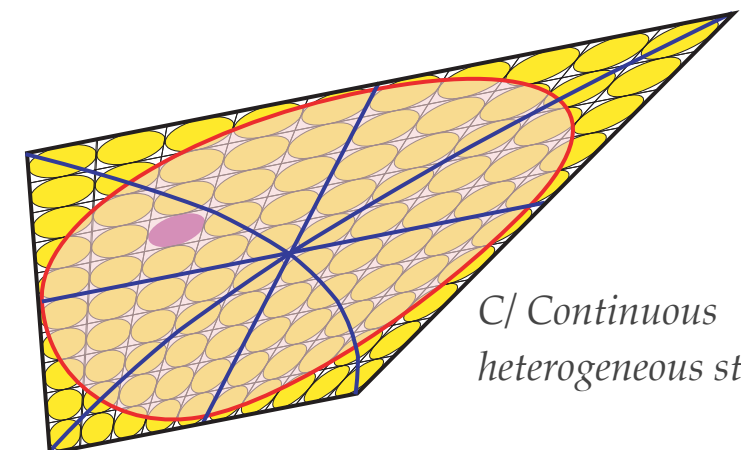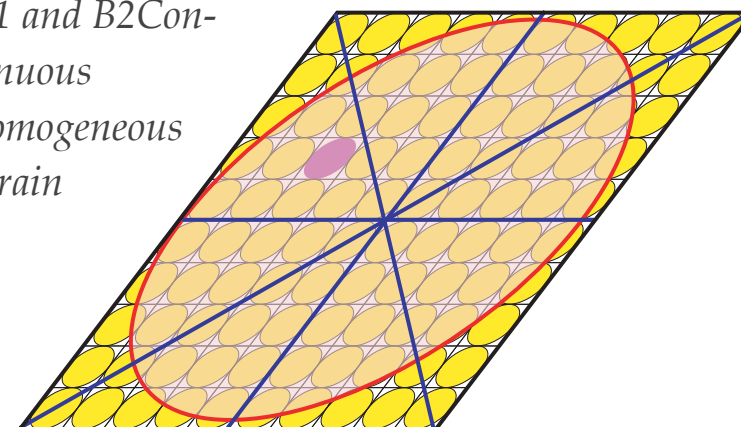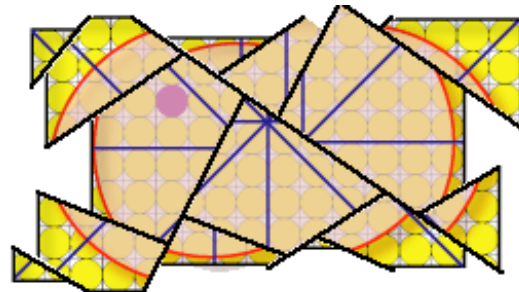
C/ Continuous heterogeneous strain

The Finite Strain Ellipsoid

When strain is heterogeneous deformation transforms a marker of spherical shape into an irregular volume, the characterisation of which is rather difficult. In constast, when strain is homogeneous, deformation transforms a spherical marker into a regular ellipsoid. This ellipsoid is called the *Finite Strain Ellipsoid* (*FSE* for short). It is characterized by 3 perpendicular axes of various length, and by convention: $\lambda_1 \geq \lambda_2 \geq \lambda_3$ (alternatively strain axes are also refered to as x≥y≥z).

This ellipsoid offers an easy way to characterise the finite strain in terms of shape (i.e. style of strain i.e. S, SL, L tectonites) and magnitude. Fortunately it is always possible to define a scale at which strain is, at first approximation, homogeneous. Finding this scale may require zooming to look closely at smaller volume, or alternatively looking from a distance at a larger volume or region.

A regular ellipsoid is fully characterized by two parameters K and D, which define the style of strain (shape of ellipsoid) and the amount of strain (i.e. how far it is from the initial perfect sphere) respectively. As shown on the right these two parameters are both functions of the ratio λ_1/λ_2 and λ_2/λ_3.

Importantly, K and D do not demand knowledge of the radius of the initial sphere, only knowledge of the principal axes of the finite strain ellipsoid.

$$\lambda_1 \geq \lambda_2 \geq \lambda_3$$

$$K = \frac{ln\,[\,\lambda_1 / \lambda_2\,]}{ln\,[\,\lambda_2 / \lambda_3\,]}$$
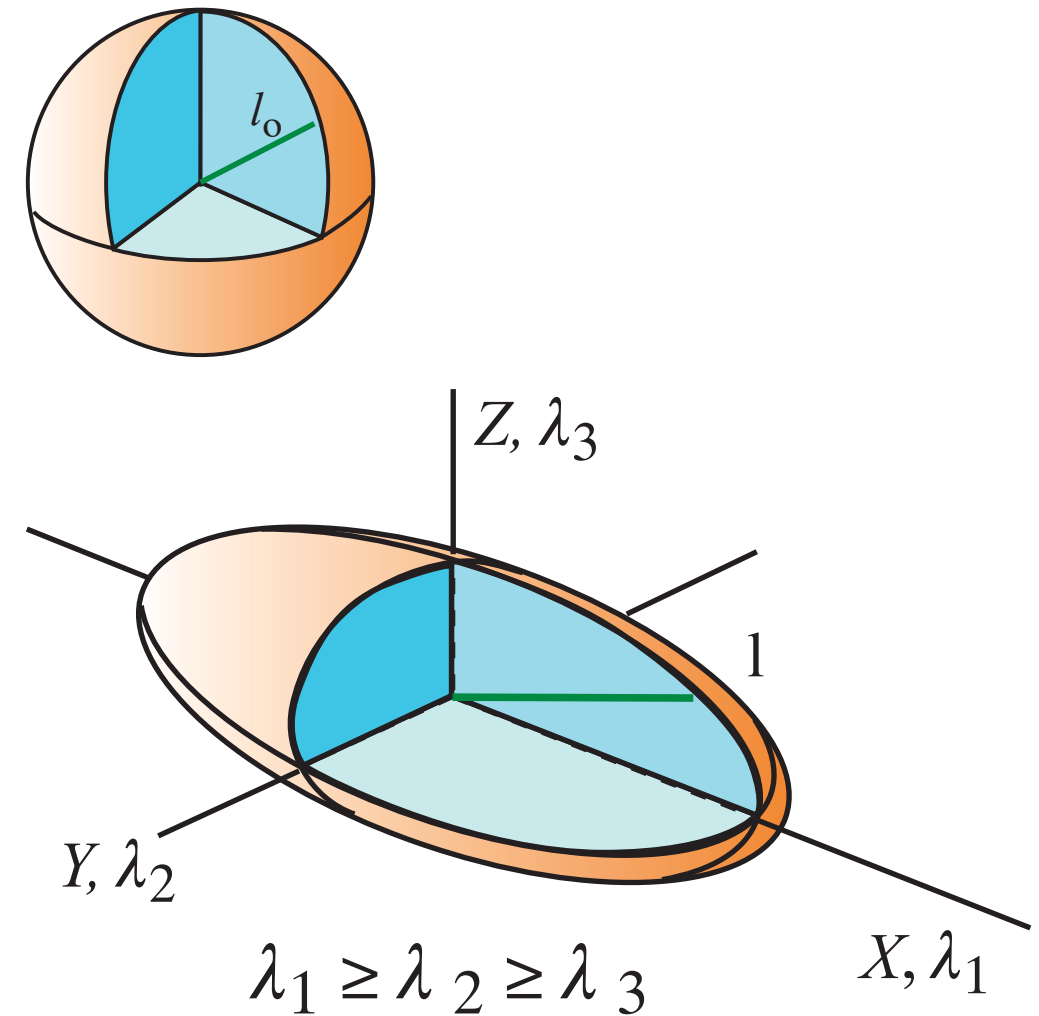
$$D = \sqrt{[\,ln\,[\,\lambda_1/\lambda_2\,]\,]^2 + [\,ln\,[\,\lambda_2/\lambda_3\,]\,]^2}$$

The Flinn's Diagram

The Flinn's diagram allows the representation of every type of regular ellipsoid on a simple 2D diagram. In this diagram, the ratio λ_1/λ_2 is the ordinate axis whereas the ratio λ_2/λ_3 is the abscissa axis. The logarithm function conveniently scales the strain axes ratios. In the Flinn's diagram, the parameter D is the distance between the origin of the graph (that represents the initial sphere) and the ellipsoid (data in red on the diagram). The parameter K is the slope of the line joining the ellipsoid and the origin of the diagram. Cigar shaped ellipsoids position themselves along the ordinate axis (i.e. K is infinite) where the strain is said to be *constrictional*. Pancake shaped ellipsoids are found along the abscissa axis ($K=0$) where the strain is said to be *flattening*. *Plane strain* ($K=1$) characterises the ellipsoids for which λ_2 axis remains constant despite strain. Pure shear leads to either constrictional or flattening strain, whereas simple shear leads to plane strain.

Strain-related fabric depends on the shape of the finite strain ellipsoid. Pancake shaped ellipsoids lead to *S tectonites* (strong schistosity, no lineation), cigar shaped ellipsoids lead to *L tectonites* (strong lineation, no schistosity). *L=S tectonites* are produced by plane strain. The various types of tectonites (or various type of fabrics) can be mapped in the Flinn's diagram (yellow inset).
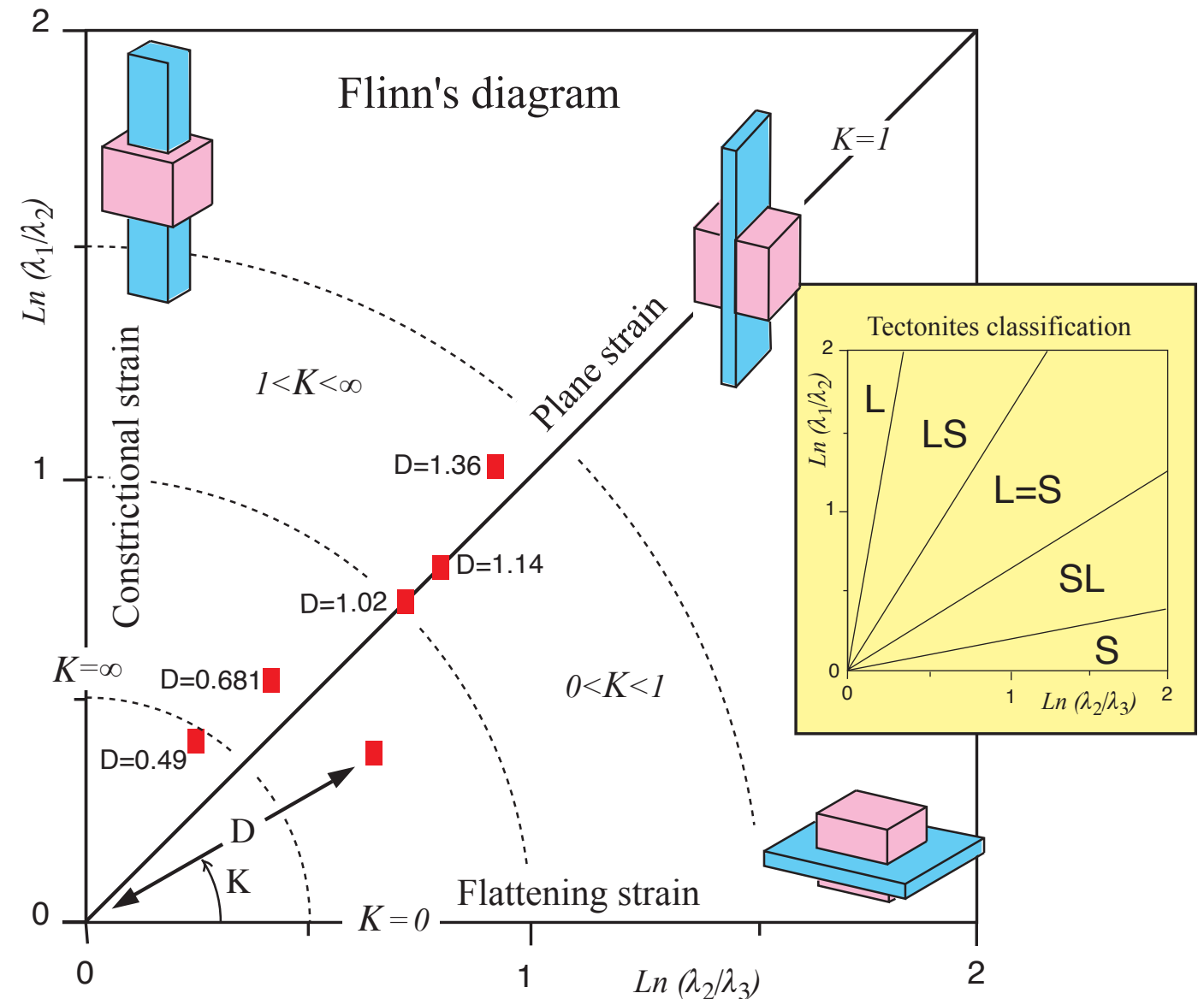
Techniques of Strain Analysis

Structural geologists have invented various methods to characterise FSE from through the analysis of rock fabrics and textures.

The Fry Method

This is one of the most elegant and easy methods to determine strain in rocks including particles initially randomly distributed in a matrix. The particle can be anything: quartz grains and quartz aggregates in rhyolites, pre-strain metamorphic minerals such as garnet in metamorphic rocks, pebbles in matrix supported pebbly sandstones, etc. The main assumption is that before deformation the spacing of these particles should be statistically isotropic i.e. in average the distance between two particles does not depend on the direction in which this distance is measured. One way to demonstrate this is to plot on tracing paper (in yellow on the sketch below) the position of particles (red circles) neighbouring a chosen particle (crossed white spot) and to repeat this operation for as many particles as possible. The result is shown on the sketch below on the right. This sketch shows that particles are disbributed around a circle the radius of which is the average distance between particles. In this case there is no strain as the distance between particles shows no directional variation.
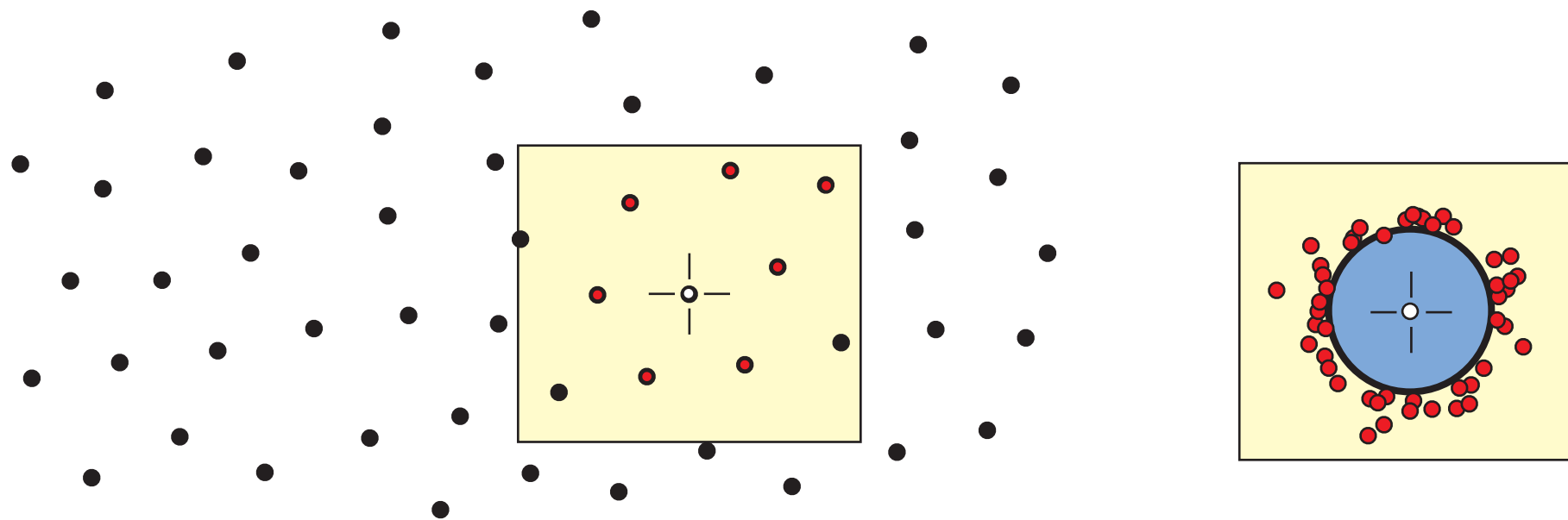
The Fry Method cont.

When strain is homogeneous, we know that spherical markers are transformed into ellipsoids. As a result the distance between the particles is changed. After deformation the distance between two particles aligned with λ_1 increases whereas two particles aligned in the direction of λ_3 get closer. To determine the three axes of this ellipsoid structural geologists work successively on two perpendicular sections (thin sections, saw cuts through hand specimens, photographs) containing two of the three principal strain axes, typically λ_1-λ_2 (i.e. the foliation plane) and λ_2-λ_3 (the plane perpendicular to the foliation and parallel to the streching lineation). The distance between a given particle (cross-hair below) and its neighboring particles is plotted on a piece of tracing paper. The tracing paper and the cross is displaced successively over as many particles as possible and their closer neighbors are plotted. The sketches on the left below show a 2D section through a deformed rock. At first glance the particles seem to be randomly distributed. The Fry analysis reveals that in reality the sample has been deformed quite significantly since the neighboring particles are distributed around an ellipse. The ratio of the long axis over short axis is a measure of the strain intensity. By plotting on a Flinn diagram the results obtained from two perpendicular sections, each containing two of the three principal axes one can fully characterise the 3D finite strain ellipsoid.
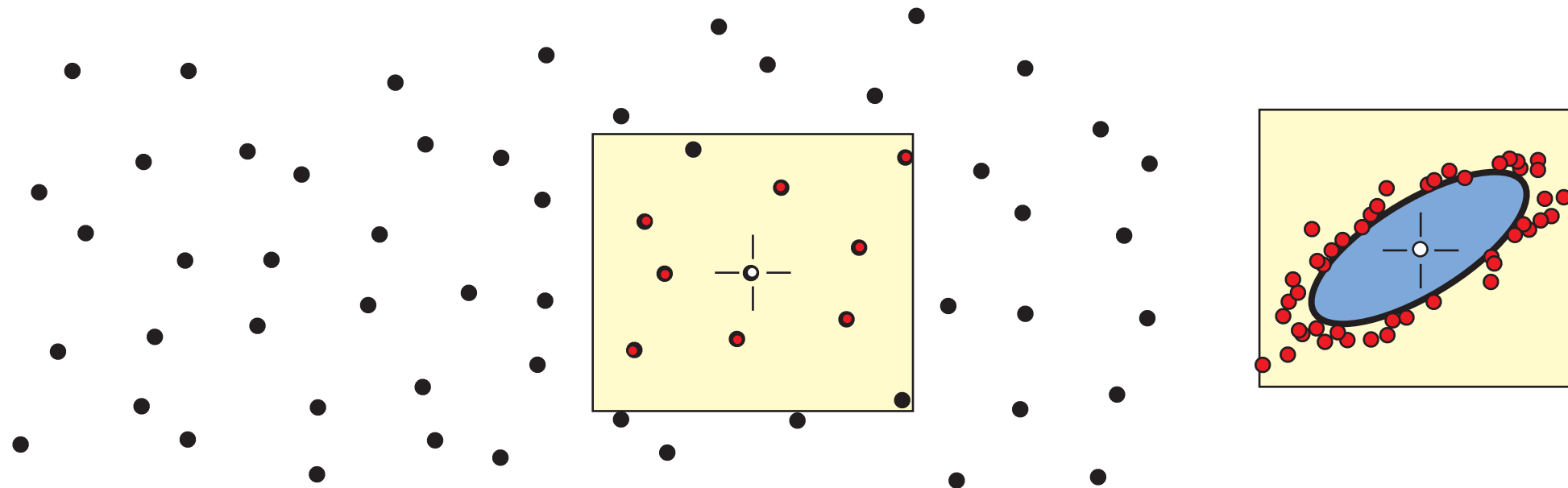
The Rf /Φ (Rfi) Method

Strain intensity can be assessed by the shape of deformable sub-spherical objects. Quartz aggregates in granitic rocks, pebbles in conglomerates, and gas vacuoles in basaltic rocks all form with a sub-spherical shape. Following homogeneous strain these spherical objects are changed into ellipsoids. By measuring the axial ratios of those ellipsoids and plotting them on a Flinn diagram, it is possible to determine the shape of the strain and its intensity. This method can only underestimate the finite strain as stronger objects record only a fraction of the total strain recorded by the bulk of the rock.

From a 2D section through a population of ellipsoidal objects of axial ratio *Rf* (sketch on the right), the *Rf /Φ* method allows the evaluation of: 1) *Ri* : the axial ratio of the objects before deformation, and 2) *Rd* : the axial ratio of the finite strain ellipse. This method is extremely useful when the initial shape of the strained objects is an ellipsoid.
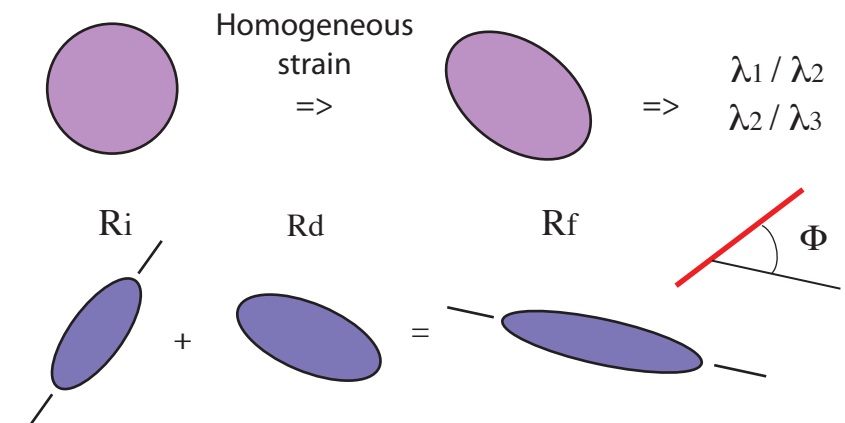
$$Rd^2 = Rfmax \cdot Rf\,min$$

$$Ri^2 = Rfmax / Rf\,min$$

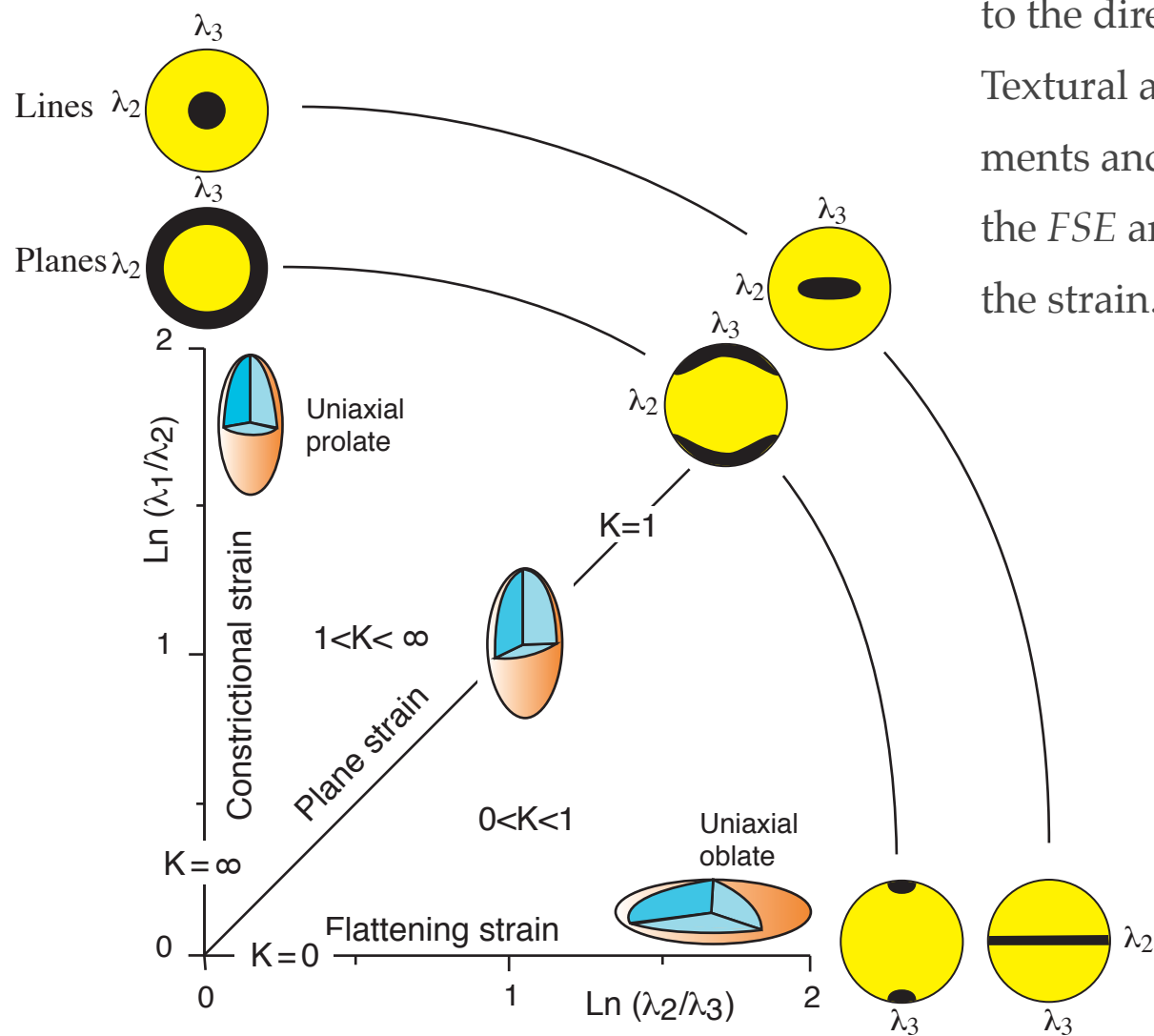$$Rd = \lambda i / \lambda j \qquad i \neq j$$
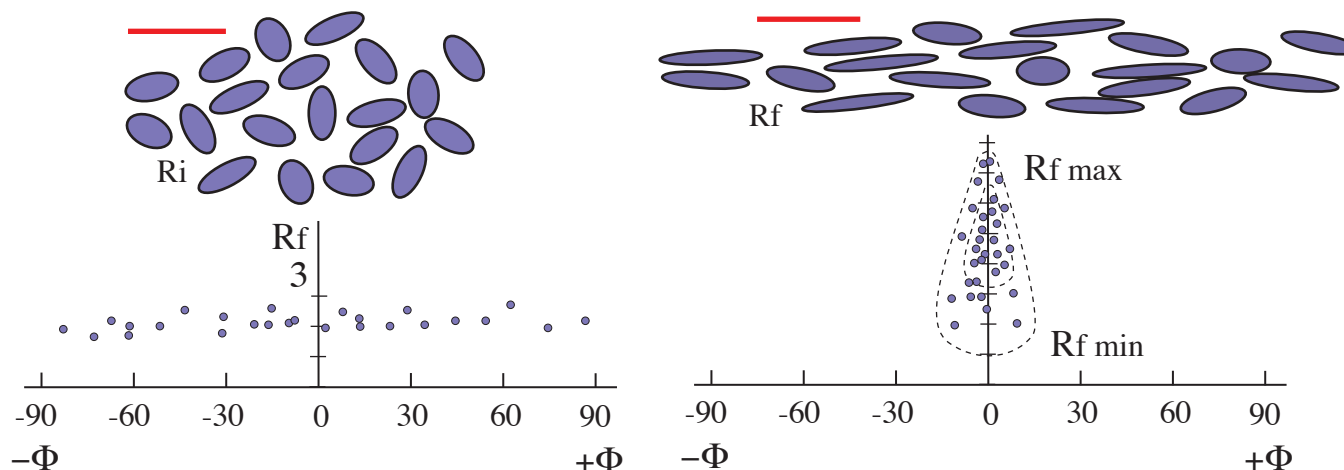
In this method, the axial ratio of the particles is plotted against *Φ*: the angle between the long axis of the particle and the direction of the average foliation (red line), or an arbitrary direction when no foliation is visible. When there is no strain *Φ* varies from +90° to -90°. When the rock is strained, the range of *Φ* decreases and the range of axial ratio of the particles (difference between *Rf max* and *Rf min*) increases. *Ri* and *Rd* are simple function of *Rf max* and *Rf min*.

Textural Analysis Method

Before deformation, planar particles such as micas are randomly distributed. On a stereonet canvas, the distribution of their poles is homogeneous. As deformation proceeds the particles rotate so their poles tend to align with the direction of maximum shortening λ_3. One can demonstrate that the orientation of planar elements after deformation is a simple function of their initial orientation and of the axial ratios (λ_1/λ_2; λ_2/λ_3; λ_1/λ_3) of the Finite Strain Ellipsoid.

A similar rotation affects linear elements which tend to become parallel to the direction of maximum lengthening (λ_1).
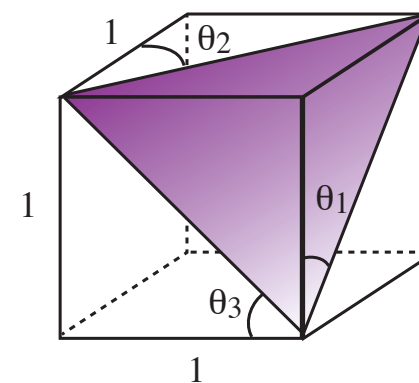
Textural analysis consists of plotting on a stereonet the pole of planar elements and linear elements. Their distribution is related to the shape of the *FSE* and the intensity of the anisotropy is related to the intensity of the strain.
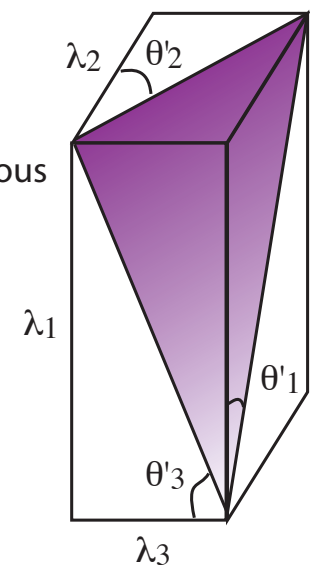

$$\text{Tan}(\theta'_1) = \lambda_3/\lambda_1 \; \text{Tan}(\theta_1)$$

$$\text{Tan}(\theta'_2) = \lambda_2/\lambda_1 \; \text{Tan}(\theta_2)$$

$$\text{Tan}(\theta'_3) = \lambda_3/\lambda_2 \; \text{Tan}(\theta_3)$$

A Little Quiz ...

1/ Define discontinuous deformation.

2/ Define continuous heterogeneous strain.

3/ How can you tell if strain is homogeneous?

4/ What happens to a sphere when it is deformed by heterogeneous strain?

5/ What is the Finite Strain Ellipsoid?

6/ What is the Flinn's diagram?

7/ Define the concept of "Plane Strain".

8/ What is an L tectonite?

9/ Describe the Fry method of strain analysis.

10/ Describe the Rf /Φ method of strain analysis.

Structural geology is the discipline concerned with the deformation of rocks and rock formations. It is a field-based discipline which aims at understanding, from observation of the landscape and the geology visible at its surface, the 3-dimensional geological architecture. This eBook aims at undergraduate students and present basic notions of structural geology including faults and fractures, folds and folds systems, boudins and boudinage, ductile shear zones, foliations and lineations, strain and strain analysis, and paleostress analysis.

Amongst all geoscience disciplines, structural geology is perhaps the one that requires the most to develop an acute sense of observation. Although digital cameras can capture in an instant and at ultra-high resolution our surrounding world, structural geologists keep drawing sketches in their notebooks. They do so for two reasons: first cameras capture the world in only 2 dimensions, yet geology can only be understood in a 3D space. Second, sketching forces careful observation and allows for the representation of the geology hidden underneath the ground surface, and that lost to erosion.

I hope this eBook, will help initiate your journey into structural geology. One of the main advantages of the eBook format is its capacity to interactively handle 3D sketches. I have designed mine using Sketchup which I highly recommend to students and colleagues.