

Introduction to Structural Geology

Workbook 1 Structural Geology - the Basics

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How to use this workbook

This workbook covers basic structural geology concepts that will be built on in later modules. It is designed as a refresher course for those who have not looked at the fundamentals of structure for a few years. By the end of this workbook you should be able to identify, describe and understand the formation of a variety of geological structures and to understand some of the basic processes that result in and occur during deformation.

Introduction to structural geology

Structural geology is the study of how rocks deform and the processes of deformation. Deformation is the change in shape, position and/or volume of an object in response to applied forces. It is closely related to the concept of strain - the permanent change in shape (in 1D, 2D or 3D) of a rock body as a result of deformation, so closely related, in fact, the two terms are often used interchangeably. However, deformation includes rigid translation and/or rotation (e.g. a fault block that moves but undergoes no internal strain) and volume change (e.g. compaction) of a rock body, whereas strain is purely the change in shape of a rock body.

Deformation is caused by forces acting on the rock body. These forces may be due to gravity (vertical force) or the movement of the tectonic plates (horizontal forces). The effect of these forces on a rock depends on the area over which they are applied: $\text{force/area}=\text{stress}$. Therefore, at its simplest, stress causes strain.

Depending on lithospheric conditions at the time of deformation, rocks may respond to stress in a brittle or ductile manner. During brittle deformation rocks fracture with strain localised along a plane whilst the rocks to either side remaining unaffected (e.g. faults and joints). During ductile deformation rocks change shape smoothly and strain is pervasive throughout the rock body (e.g. folds).

Worksheet 1: Deformation, strain and stress

Deformation: The change in shape, position and/or volume of a rock in response to applied forces. Deformation is determined by comparing the rock's deformed and undeformed states. It can be broken down into rigid body deformation (translation and rotation) and non-rigid body deformation (strain and volume change).

Rigid body deformation

Rigid body deformation occurs where a rock mass moves or rotates with no change of shape (figure 1). This is only detectable if there is an external reference frame. An example is the rotation of the African continent since the Jurassic. This rotation can be recognised from palaeomagnetic data.

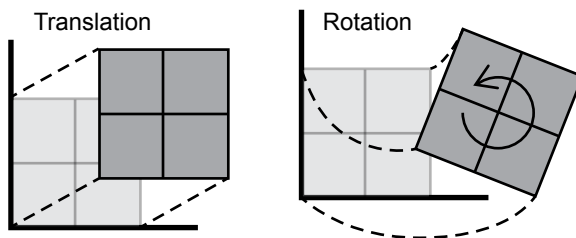


Figure 1: Rigid body deformation involves translation and/or rotation of an object.

Translation

During translation every point in a rock body undergoes the same displacement. There is no distortion and no change of shape

Rotation

Rotation is the same as translation only with a rotational component

Non-rigid body deformation

Strain

Strain is the change in shape of a rock body during deformation (figure 2).

Volume change

Volume change occurs where a rock body increases or decreases in volume. In two dimensions this is a change in area. A rock body may retain the same shape if the volume (area) change is the same in all directions or it may change shape if the volume (area) change varies. Compaction is an example of a change in shape and reduction in volume due to vertical compression.



Figure 2: A sheared trilobite. An example of non-rigid body deformation (Miller, 2012).

Homogeneous / heterogeneous deformation

Homogeneous deformation

Homogeneous deformation occurs where deformation is constant across a rock body, that is different parts of an object deform by the same amount (figure 3a). In homogeneous deformation straight lines remain straight, parallel lines remain parallel and circles deform to ellipses.

Heterogeneous deformation

Heterogeneous deformation occurs where the deformation varies across a rock body, so different parts of an object deform by different amounts (figure 3b).

Whether deformation is homogeneous or heterogeneous can depend on scale; a large area of heterogeneous deformation may be broken down into smaller areas of homogeneous deformation for analysis.

Strain parameters

The strain parameters **longitudinal strain** and **angular shear strain** are used to measure homogeneous strain.

Longitudinal strain

Longitudinal strain is strain in a single direction (1D) or the change in the length of a line. It may be measured in units (centimetres, kilometres etc) or expressed as the ratio elongation (e).

$$\text{Elongation (e)} = \frac{l_1 - l_0}{l_0}$$

where:

l_0 is the original length of the line

l_1 is the new (observed) length of the line

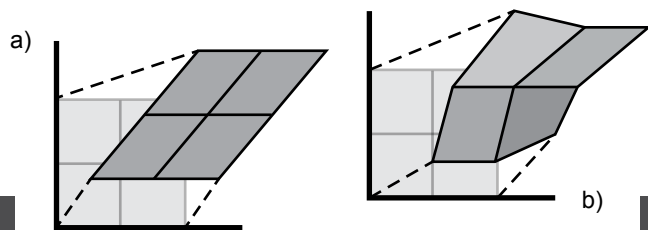


Figure 3: a) Homogeneous deformation: straight lines remain straight, parallel lines remain parallel and circles deform to ellipses. b) Heterogeneous deformation: different parts of an object deform by different amounts.



Figure 4: Folded quartz vein, an example of negative elongation (R. Butler).

Elongation is negative (-) for contraction (the line has decreased in length) (figure 4) and positive (+) for extension (the line has increased in length) (figure 5) .



Figure 5: An extended belemnite, an example of positive elongation. The dark areas are the original fossil and white a calcite infill (J. Houghton).

Longitudinal strain can be used to calculate the extension or contraction along a cross section where a single marker horizon can be measured before and after deformation (figure 6).

Angular shear strain

Angular shear strain is the strain in a plane (2D) or the change in angles. Angular shear is the deflection of an orthogonal marker. This is the change in angle between a pair of lines that were originally orthogonal. An example of an orthogonal marker would be a fossil with bilateral symmetry (the trilobite in figure 2).

Shear strain (γ) is the tangent of the change in angle between the originally orthogonal pair of lines (figure 7).

Shear strain is positive for a clockwise deflection of the marker and negative for an anticlockwise deflection.

Shear strain $\gamma = \tan \psi$

where:

γ is the shear strain

ψ is the change in angle from the perpendicular

Figure 8 shows sheared burrows (yellow lines). The burrows were originally perpendicular (white lines) to bedding (red line). The burrows are now at approximately 50° to bedding giving a shear strain of -1.2. Shear strain is negative as the burrows have rotated anticlockwise from their original position.

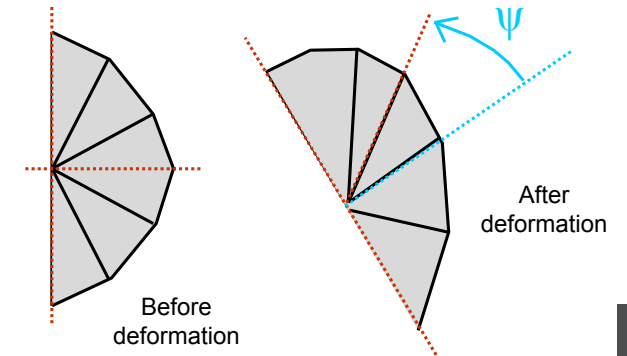


Figure 7: Angular shear of an originally orthogonal marker (e.g. fossil with bilateral symmetry). Before deformation the base and mid lines of the fossil are orthogonal. Shear across the fossil results in the deflection of these markers.

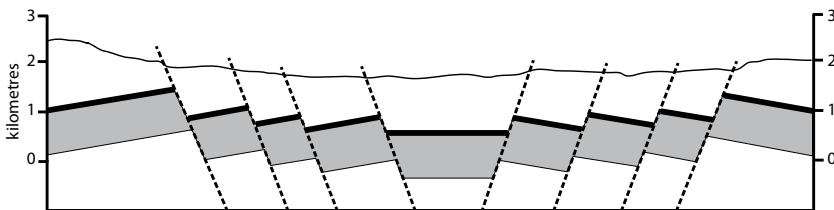


Figure 6: The black marker horizon can be used to calculate the longitudinal strain across the section.

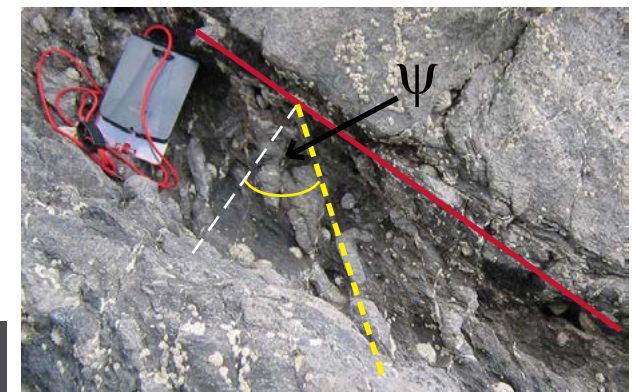


Figure 8: Sheared burrows in the limb of a syncline at Stackpole Key in Pembrokeshire (G. Lloyd). See text for details.

Strain in 2D - the strain ellipse

The strain ellipse is a method of representing the amount of strain a rock has undergone. It uses an initially circular marker that is deformed to an ellipse (figure 9). The strain ellipsoid is used for strain in three dimension.

The value of longitudinal strain depends on original orientation of a line. Lines that are parallel or close to parallel with the Z axis will contract, lines that are parallel or close to parallel with the X axis will elongated and lines that are parallel or close to parallel to the lines of no finite longitudinal strain will undergo contraction followed by elongation. The lines of no finite longitudinal strain separate the zone of elongation (yellow) from the zone of contraction (blue).

The value/sense of angular shear strain depends on original orientation of the orthogonal lines of the strain indicator. Where the lines of the strain indicator are parallel with the strain axes (X and Z) shear strain is zero. Where they are at an angular shear strain occurs (figure 10).

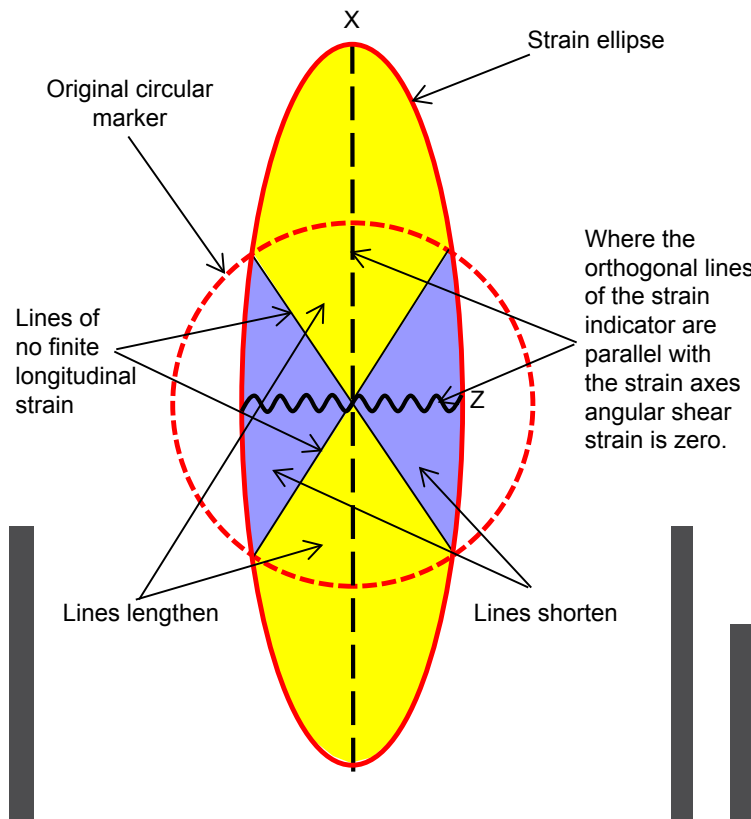


Figure 9: The strain ellipse.

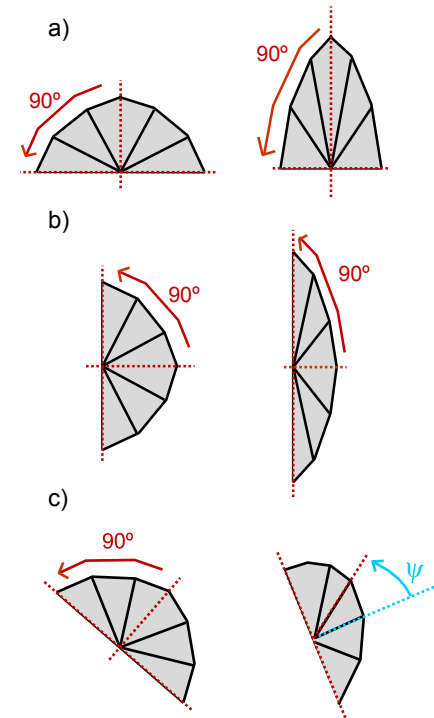
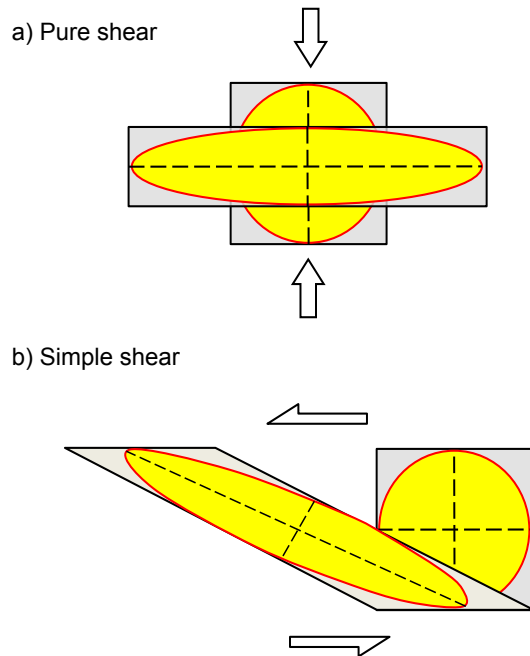


Figure 10: Fossils a) and b) have their orthogonal lines parallel with the strain axes and show no shear strain during deformation. Fossil c) is at an angle to the strain axes and shows shear strain during deformation.

Pure shear and simple shear

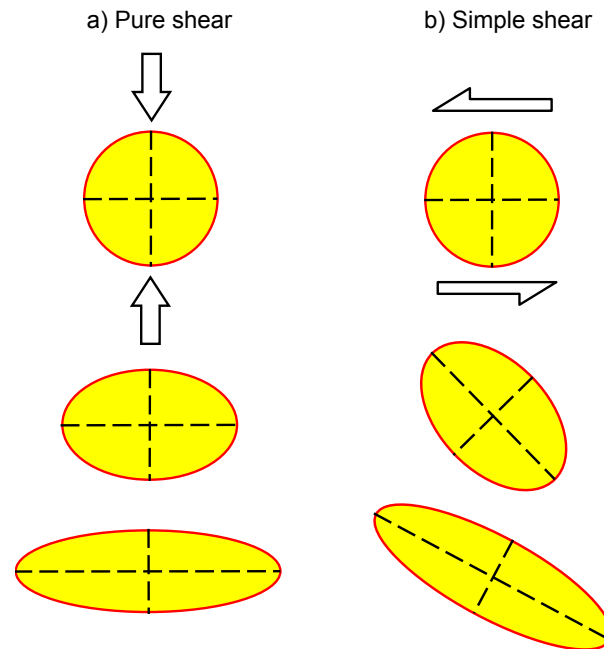
Pure shear

Pure shear is coaxial deformation. Contraction and elongation are parallel to the strain axes, so there will be no rotation of the axes from their original positions (figure 11a).



Simple shear

Simple shear is non-coaxial deformation (figure 11b). During deformation the strain axes rotate. The amount of rotation depends on amount of strain; the greater the strain, the greater the rotation.



Pure and simple shear are end members of a range of potential deformations within a plane. An object can have the same initial shape and the same deformed shape (see figure 12) but its deformation history – the path the rock took from its original to its final shape may be different depending upon whether it underwent pure or simple shear or some combination thereof.

Stress

Stress is defined as a pair of equal and opposite forces acting on a unit area of a rock body.

$$\text{STRESS} = \text{Force/Area}$$

Figure 13 illustrates the effects of the same force applied over different areas.

On a large scale, movement of the plates and gravity create the forces that act on rock layers, with gravity acting vertically and plate movements horizontally. On a smaller scale the picture

Figure 11: Pure shear and simple shear of an originally circular object.

Figure 12: Comparison of the development of a) pure and b) simple shear strain ellipses.

is more complicated with local stress fields created by the effects of topography, pre-existing structures, burial, uplift, thermal effects from

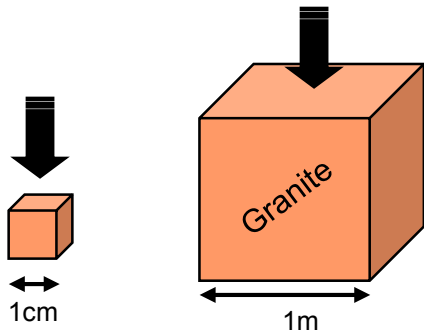


Figure 13: When hit with the same force the small cube is more likely to shatter than the large cube, illustrating the effects of the area over which a force is applied.

intrusions, contacts between rocks of different properties, anisotropic rocks etc.

Stress on a surface such as bedding can be broken down into normal stress σ_n (sigma n) which acts perpendicular to the surface and shear stress σ_s (sigma s) oriented parallel to the surface (figure 14).

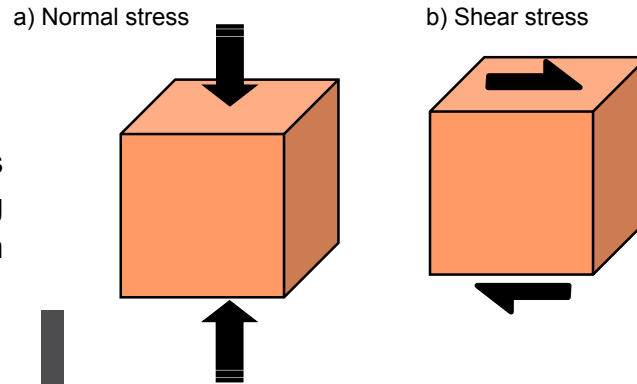


Figure 14: Normal stress (σ_n) acts perpendicular to a surface. Shear stress (σ_s) act parallel to a surface.

The stress field acting on any particular point (the local state of stress) can be considered by looking at the normal stresses acting on the point. These are known as the principal stress and they are orthogonal to each other.

σ_1 = direction of greatest compression

σ_2 = intermediate between σ_1 and σ_3

σ_3 = direction of least compression or greatest extension

Hydrostatic stress occurs where the principal stresses have the same value: $\sigma_1 = \sigma_2 = \sigma_3$. This is the state of stress found in fluids and does not cause strain, although it may cause a loss in volume. For strain to occur a **deviatoric stress** state where $\sigma_1 > \sigma_2 > \sigma_3$ is needed (figure 15).

Whilst there may be a generalized agreement between the orientation of the principal stresses and the orientation of the axes of the strain ellipse localized stress fields mean a straight forward correlation is often not possible.

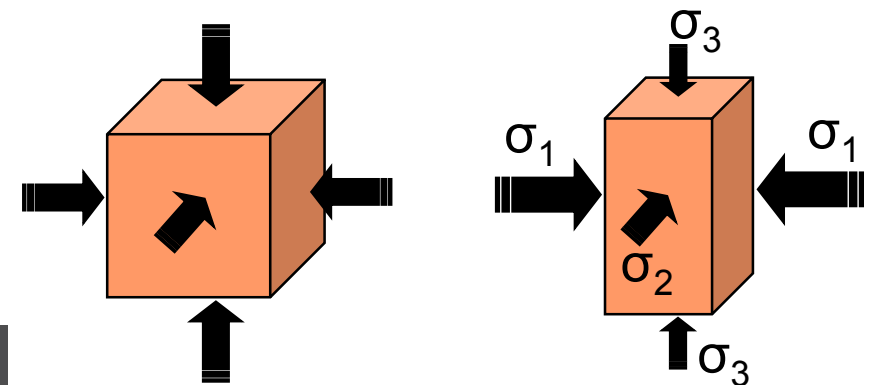


Figure 15: For deformation to occur a deviatoric stress state is needed

Worksheet 2: Brittle deformation

Brittle deformation occurs where the stress exceeds local rupture strength of the rock and the rock fractures. Deformation is localised along a plane with the rocks to either side of the fracture unaffected. It occurs in lower temperatures and pressures of the upper crust. Joints, fissures and faults are all examples of brittle deformation.

Faults

A fault is a surface or narrow zone with parallel displacement and offset of layers either side of the fracture (figure 16).

Footwall: Fault block beneath the fault surface.

Footwall cut-off: Where a layer or feature in the footwall is cut by the fault.

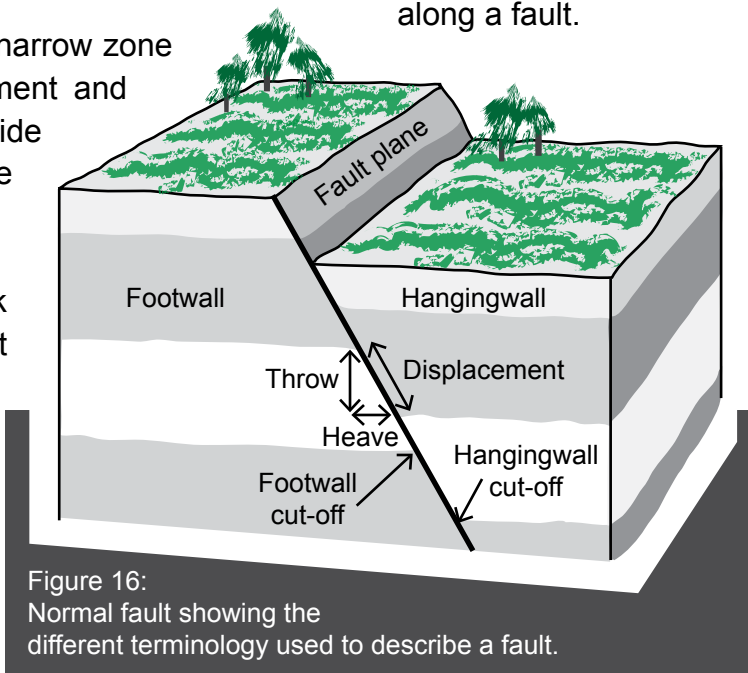


Figure 16: Normal fault showing the different terminology used to describe a fault.

Hangingwall: Fault block above the fault surface.

Hangingwall cut-off: Where a layer or feature in the hanging-wall is cut by the fault.

Displacement: The relative offset of points once adjacent on either side of a fault.

Throw: Vertical component of displacement along a fault.

Heave: Horizontal component of displacement along a fault.

Fault types

Dip-slip: Movement parallel to dip of the fault plane. Both normal, thrust and reverse faults are dip-slip faults.

Normal fault: Hangingwall moves down relative to the footwall.

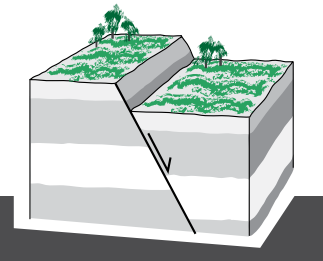


Figure 17a: Normal fault

Thrust fault: Hangingwall moves up relative to the footwall. A reverse fault is a high angle thrust fault.

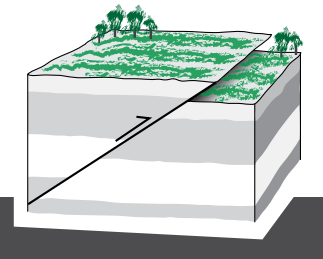


Figure 17b: Thrust fault

Strike-slip: Movement parallel to the strike of the fault plane.

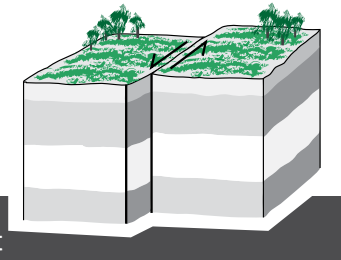


Figure 17c: Strike-slip fault

Oblique-slip faults are a combination of dip-slip and strike-slip movement where the hangingwall moves obliquely with respect to the footwall.

Normal faults

A normal fault is a fault where the hangingwall moves down relative to the footwall. It usually dips at about 50 – 60°, although local conditions can cause this to vary. Sigma one is vertical and sigma three is horizontal (Figure 18).

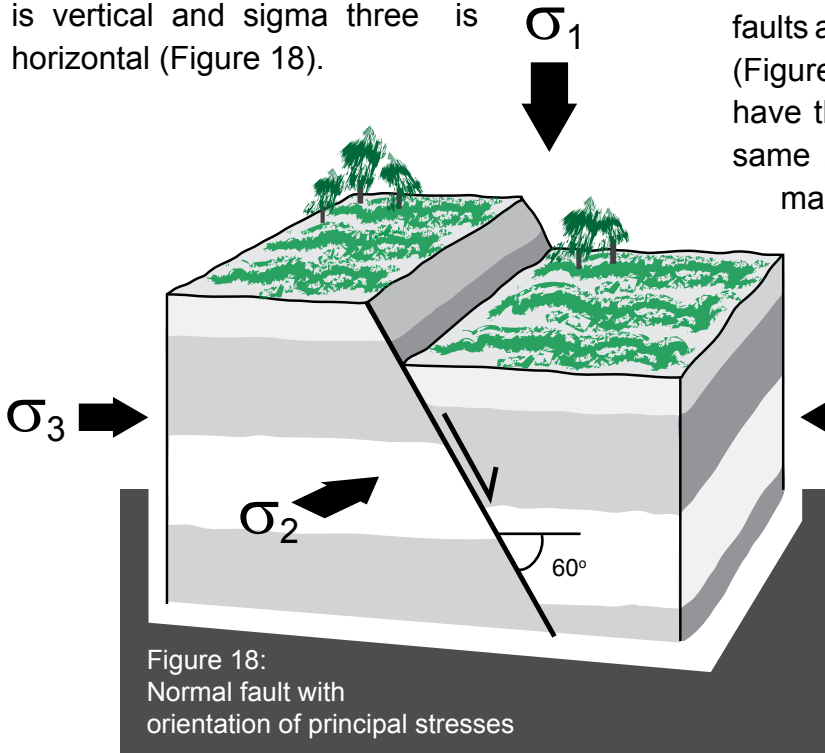


Figure 18: Normal fault with orientation of principal stresses

Listric faults are normal faults that decrease in dip with depth (Figure 19 and 21). The hangingwall is folded into a roll-over anticline. Listric faults often occur at a kilometre scale.

Domino faults are a series of parallel normal faults and fault blocks that rotate during extension (Figure 20 and 22). In the idealized model, faults have the same dip and the fault blocks are the same size. The fault blocks behave in a rigid manner and rotate at the same time and at the same rate. This style of faulting leads to space problems at the base of the fault system which are usually accommodated by ductile deformation along a weak layer such as salt or a thick series of shales.

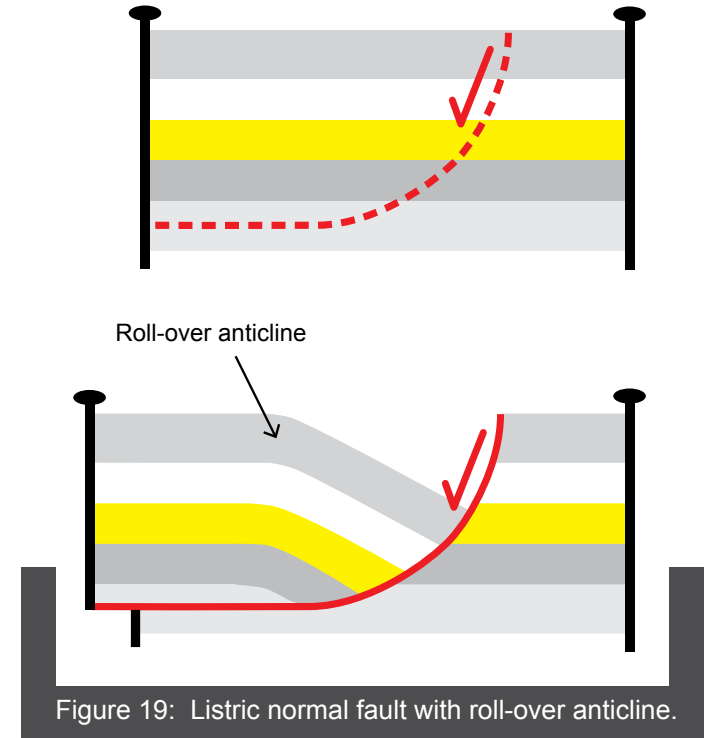


Figure 19: Listric normal fault with roll-over anticline.

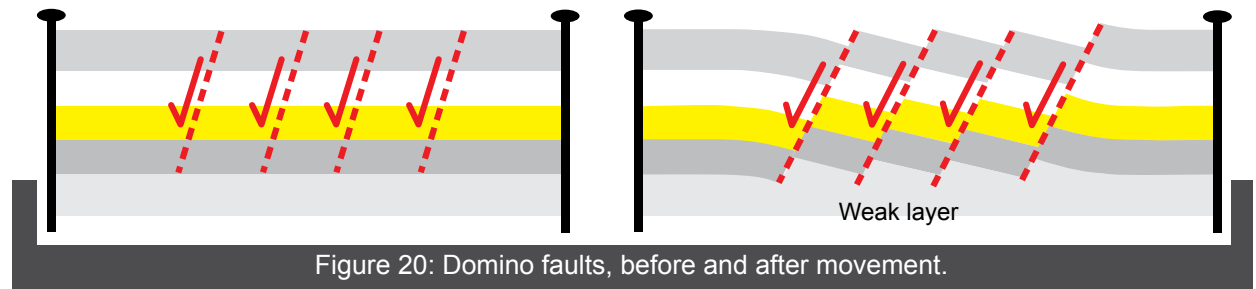


Figure 20: Domino faults, before and after movement.



Figure 21: Normal faults in a quarry face. Note the small listric fault in the top right hand corner (RDR).



Figure 22: Domino faults in lavas (Afar Rift Consortium 2012).

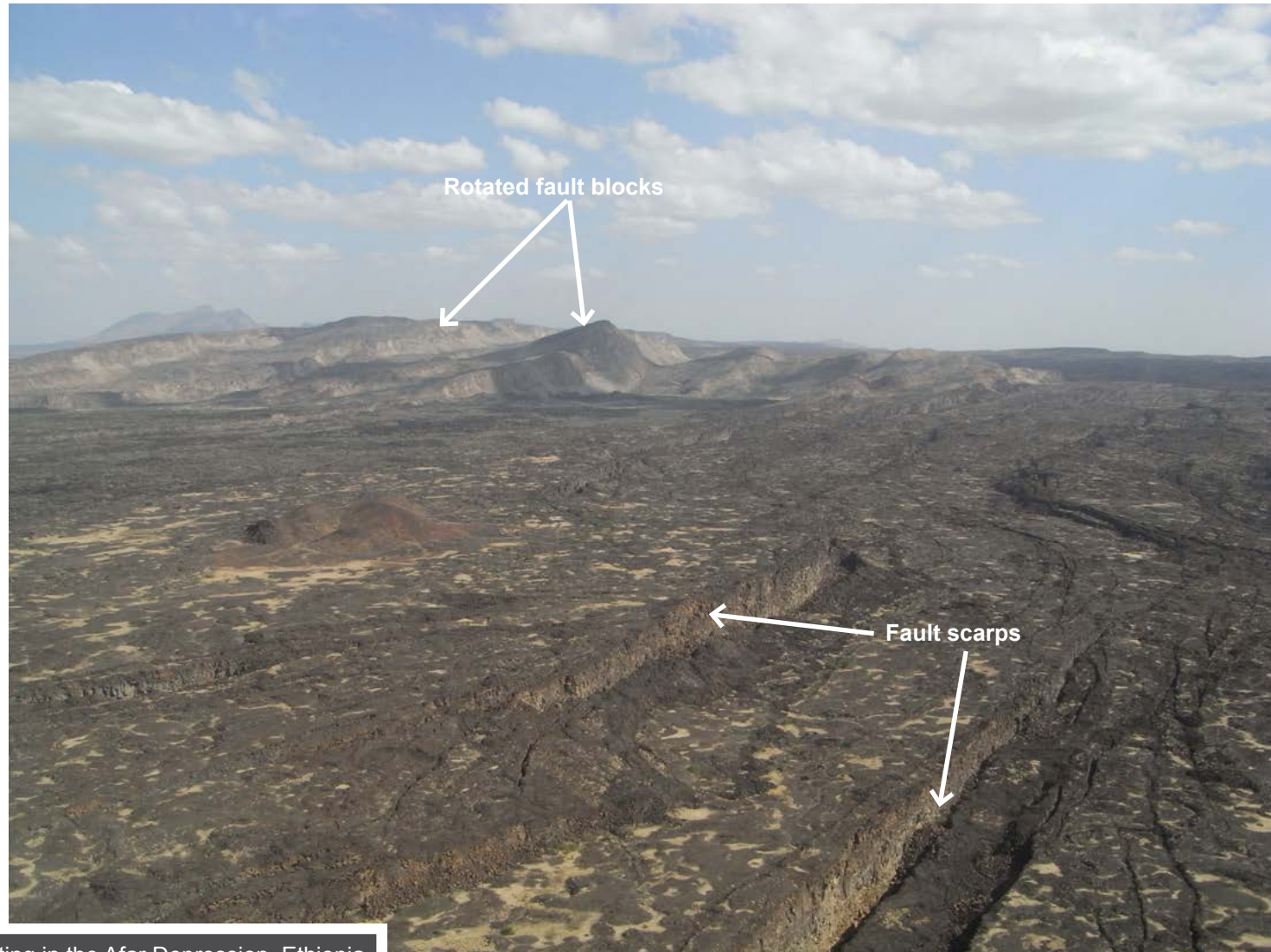


Figure 23: Rifting in the Afar Depression, Ethiopia. Fault scarps in the foreground and tilted fault blocks in the background (Afar Rift Consortium, 2012).

Thrust faults

A thrust fault is one where the hangingwall moves up relative to the footwall (figure 24). It usually dips at about 0 – 30°, although local conditions can cause this to vary. Sigma one is horizontal and sigma three is vertical. As the hangingwall moves up the footwall older rocks are placed on top of younger rocks.

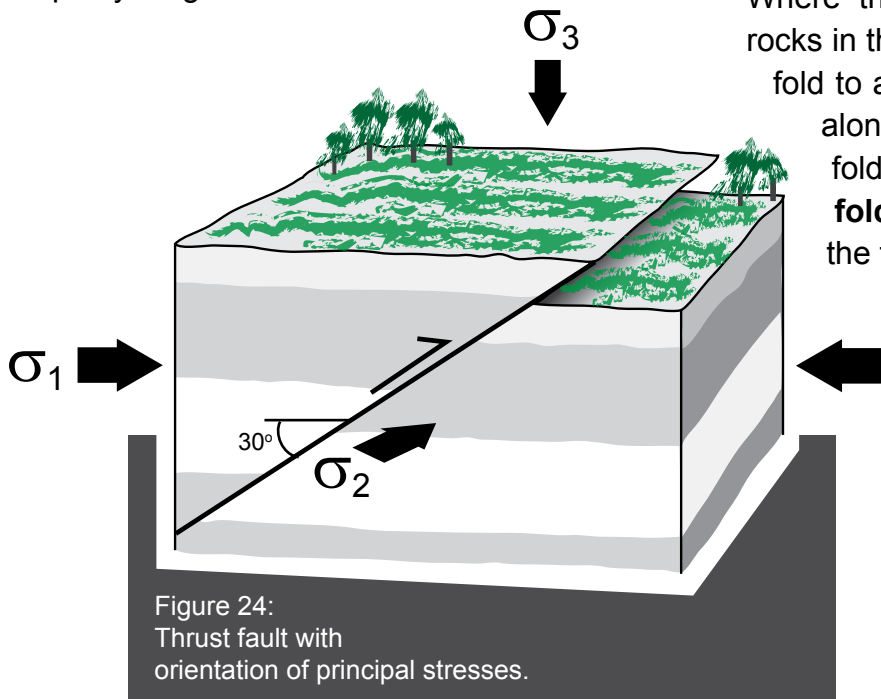


Figure 24:
Thrust fault with
orientation of principal stresses.

Thrust often run parallel to bedding (**flat**) then cut up through the layers (**ramp**) before finding another weak layer to follow (figure 25). The footwall and hangingwall are considered separately when applying the terms ramp and flat.

Where the thrust cuts upwards, the rocks in the hanging-wall are forced to fold to accommodate the shortening along the thrust. The resulting fold is known as a **fault bend fold** (figure 29). The layers in the footwall remain underformed.

Duplexes develop as displacement ceases up a ramp and the thrust cuts through into the footwall to create a new ramp. New thrusts form progressively in the footwall and older thrusts are rotated and folded by younger ones. This results in a well-ordered geometry with predictive elements (figure 26 and 27).

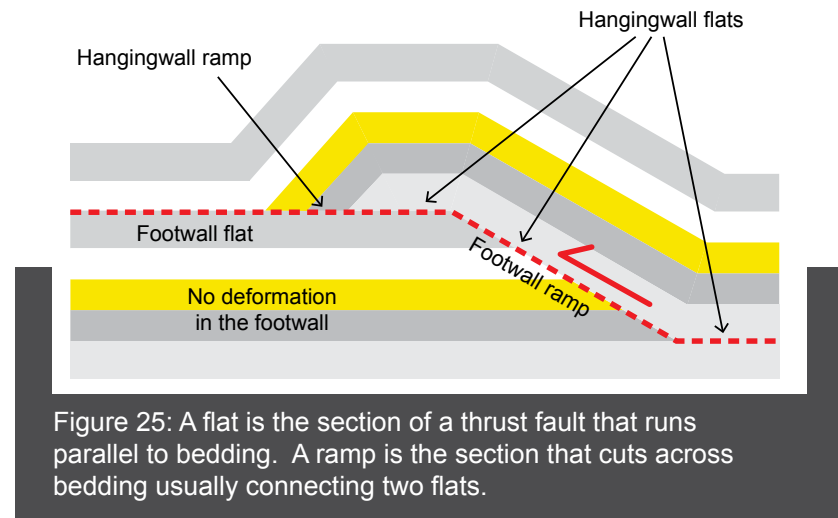


Figure 25: A flat is the section of a thrust fault that runs parallel to bedding. A ramp is the section that cuts across bedding usually connecting two flats.

Fold and thrust belts

Figure 28 shows the classic fold and thrust geometry found in the forelands of mountain belts.

Thrust tip: Leading edge of a thrust fault where displacement ceases.

Blind thrust: The foremost thrust of a sequence, which dies out before it reaches the surface.

Imbricate fan: A series of thrusts extending from a floor thrust to the surface.

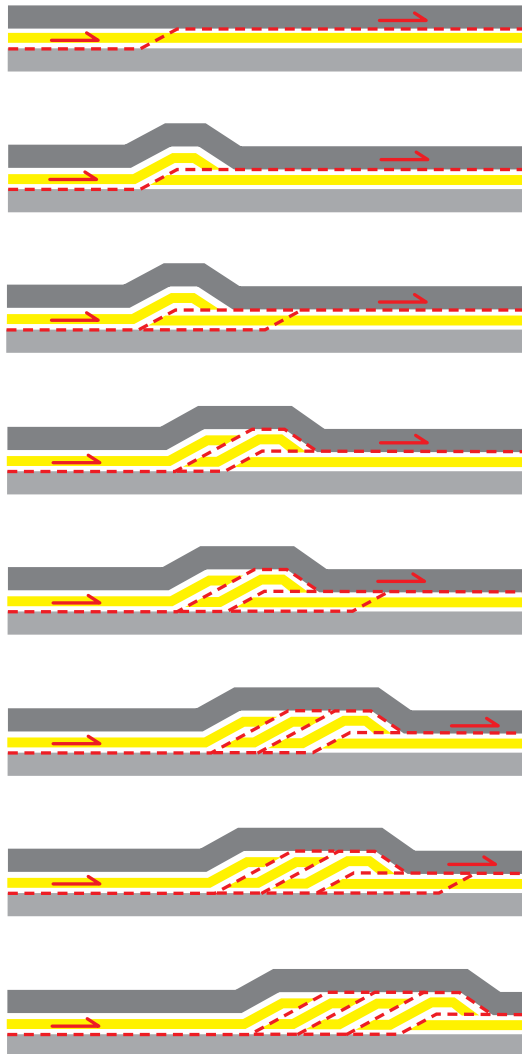


Figure 26: Formation of a duplex

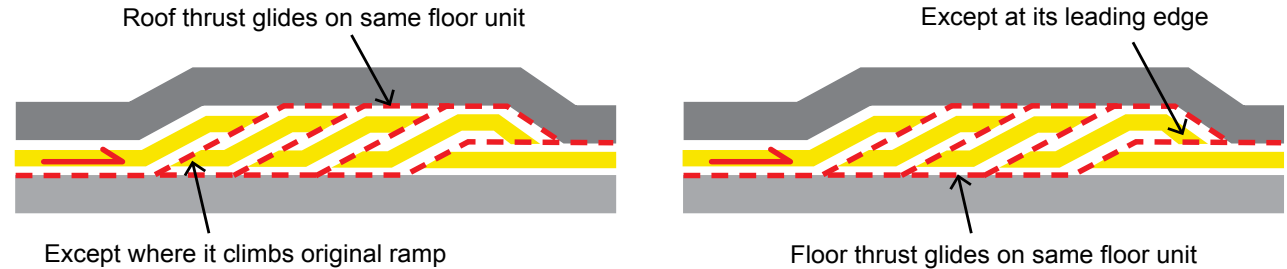


Figure 27: The predictive geometry that results from the formation of a duplex. The roof thrust glides through the upper white unit except where it climbs the original ramp and the floor thrust glides through the lower white unit except at its leading edge.

Imbricates: A series of thrusts with the same dip and branching from the same floor thrust, may be part of a duplex or imbricate fan.

Branch lines: Where a thrust branches away from or onto a floor or roof thrust.

Horse: A fault bound block within a duplex.

Roof thrust: Uppermost thrust of a duplex.

Floor thrust: Basal thrust of a duplex or imbricate fan, also known as a sole thrust.

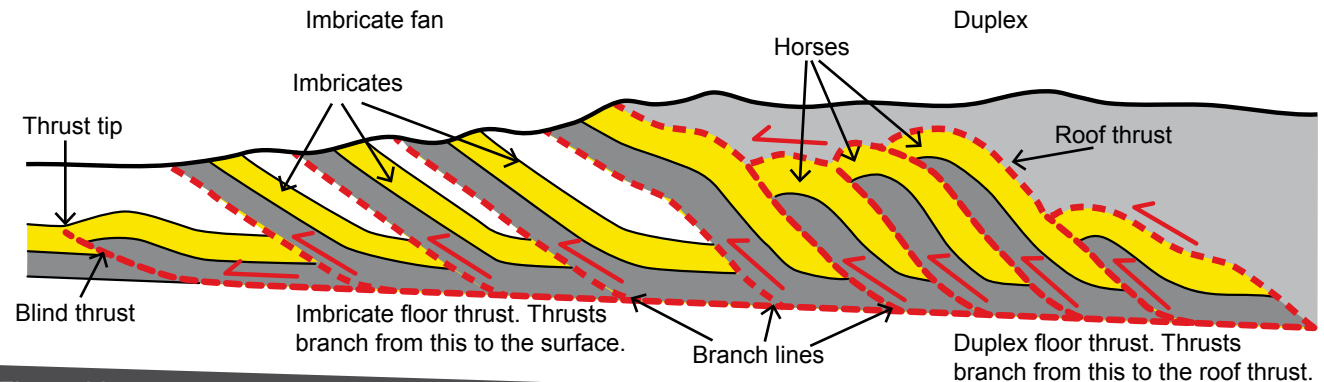


Figure 28: Fold and thrust belt. For definition of terms see text.



Figure 29a: Thrust fault with a fault bend fold in the hangingwall and undeformed footwall (Miller 2012).

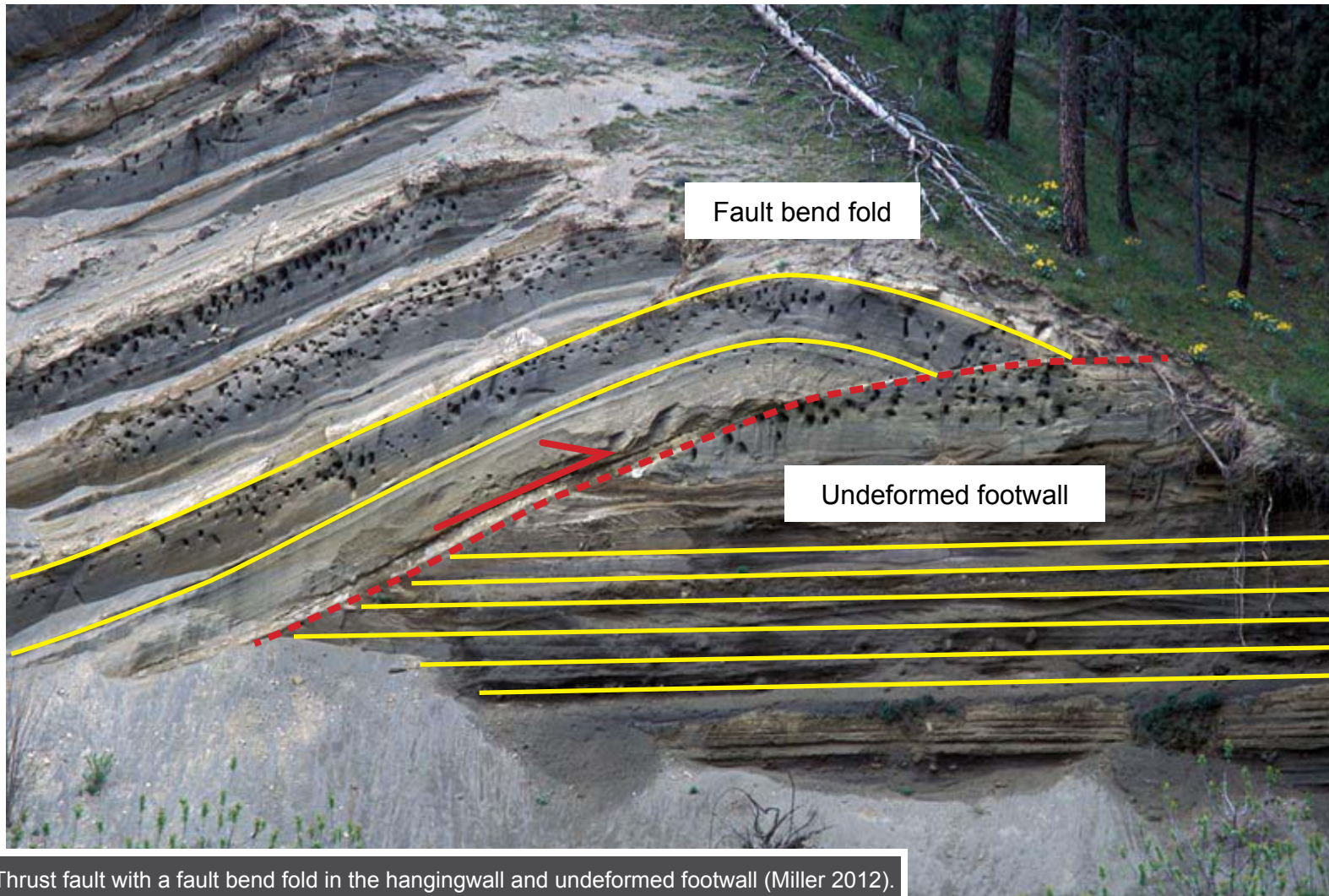


Figure 29b: Thrust fault with a fault bend fold in the hangingwall and undeformed footwall (Miller 2012).

Strike-slip faults

A strike-slip fault is one where the fault blocks move horizontally past each other. Sigma one and three are both horizontal and sigma two is vertical (figure 30). Where relative displacement is to the right the fault is dextral, where it is to the left the fault is sinistral (figure 31 and 33). Strike-slip faults are usually vertical or close to vertical.

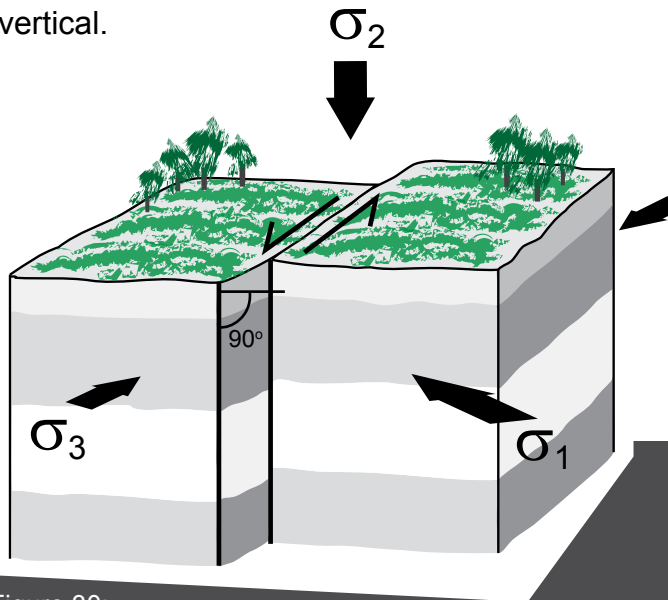


Figure 30: Strike-slip fault with orientation of principal stresses.

Bends in strike-slip faults result in areas of compression or extension depending on the direction of movement relative to the bend.

Where the bend is in the same direction as movement, an area of extension develops and a pull-apart basin bound by the strike-slip fault and normal faults forms. This is known as a releasing bend (figure 32a and 34).

Where the bend is in the opposite direction to movement an area of compression develops with localised thrusting and folding. This is known as a restraining bend (figure 32b and 35).

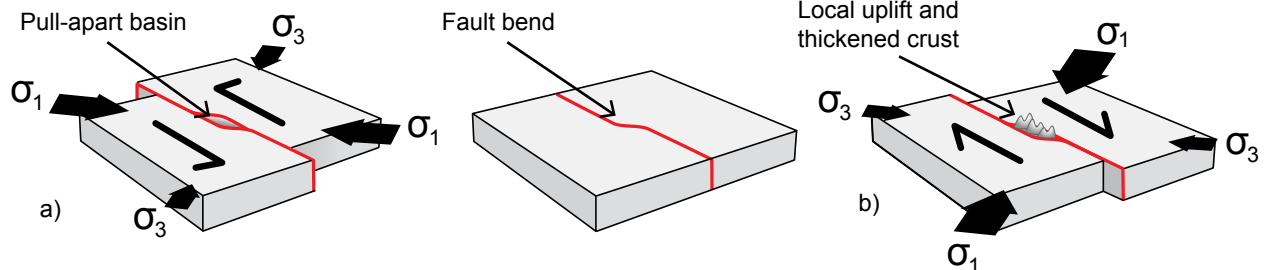


Figure 32: a) Formation of a releasing bend. b) Formation of a restraining bend.

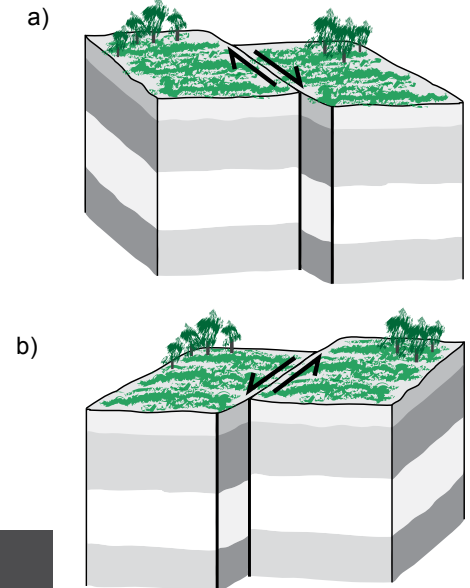


Figure 31: a) Dextral displacement: Fault blocks move to the right relative to each other. b) Sinistral displacement: Fault blocks move to the left relative to each other.

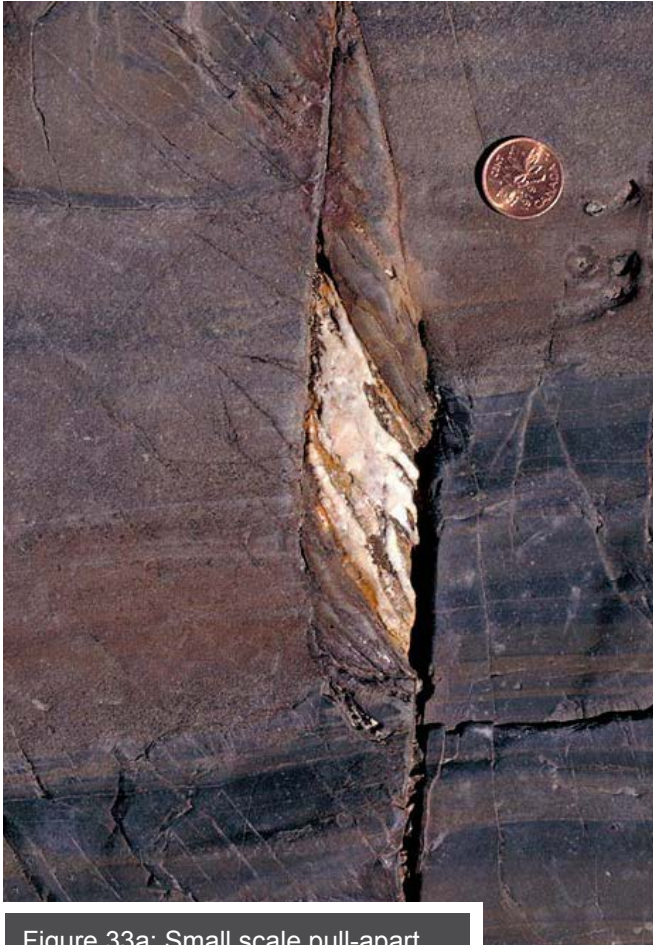


Figure 33a: Small scale pull-apart filled with calcite fibres on sinistral strike-slip faults (Miller, 2012).

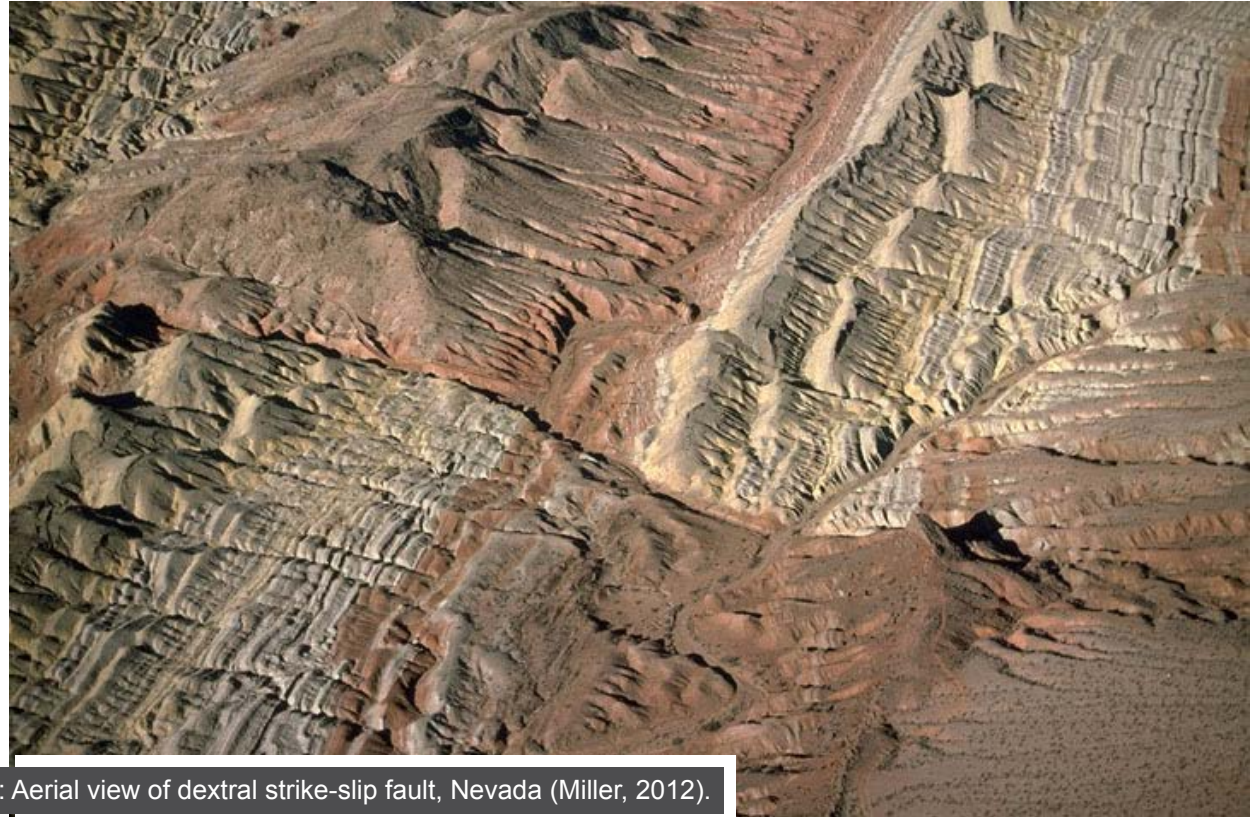


Figure 34a: Aerial view of dextral strike-slip fault, Nevada (Miller, 2012).



Figure 35: Metre-scale folds associated with a restraining bend along a strike-slip fault on the Northumberland Coast (J.Houghton)



Figure 33b: Small scale pull-apart filled with calcite fibres on sinistral strike-slip faults (Miller, 2012).



Figure 34b: Aerial view of dextral strike-slip fault, Nevada (Miller, 2012).



Figure 35: Metre-scale folds associated with a restraining bend along a strike-slip fault on the Northumberland Coast (J.Houghton)

Worksheet 3: Ductile deformation

Ductile deformation occurs where rocks change shape smoothly, without breaking, in response to stress. This shape change is pervasive – it affects all of the rock. Large scale ductile deformation occurs at higher temperatures and pressures and so deeper in the crust than brittle deformation. However, ductile deformation can also occur at shallower levels in the crust in association with the formation of faults or in weaker rocks (e.g.

salt). Folds, cleavage, boudinage and shear zones are all examples of ductile deformation.

Folds

Folds can form on all scales and under a variety of conditions. Active folding is the response of layers of different competence to layer parallel compression. Passive folding occurs where

layering has no mechanical influence on the folds formed, such as when layers fold in response to movement along a fault. This section looks at folds formed by active folding, but much also applies to folds related to faulting.

Fold terminology (figure 37)

Anticline: A fold with older rocks in its core.

Antiform: A fold where the limbs close upwards.

Axial plane: A plane that connects all the hinge lines through a fold, also known as the axial surface (3D term).

Axial trace: Where only an edge of the axial plane is seen, e.g. when a fold is seen in cross section or on a map. The hinge line runs along the top of the axial plane and so is also an axial trace (2D term).

Facing: The direction of younging along the fold axial plane. In an upward facing fold the beds are the right way up and get younger from the bottom to the top of the axial plane.

Hinge: Point of maximum curvature on a fold (2D term).

Hinge line: Line of maximum curvature along a fold (3D term).

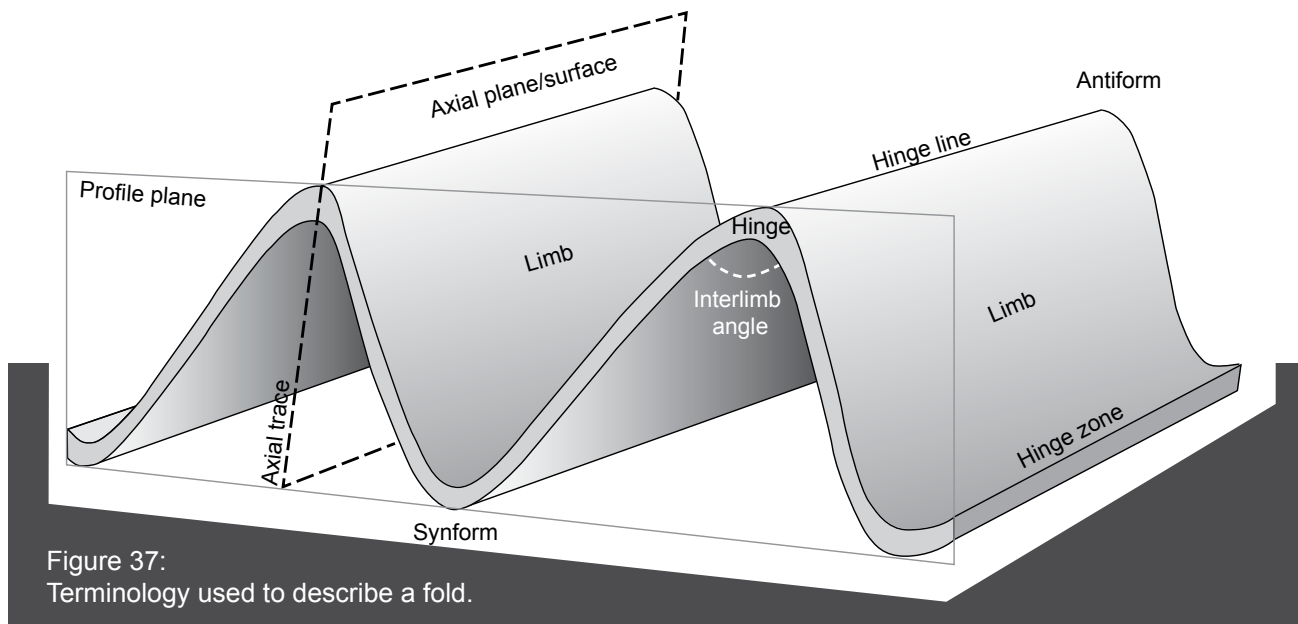


Figure 37:
Terminology used to describe a fold.

Interlimb angle: The angle between the two limbs of a fold.

Limb: Bed segments between hinges.

Profile plane: The true cross section through a fold, perpendicular to the axial plane.

Syncline: A fold with younger rocks at in the core.

Synform: A fold where the limbs close downwards.

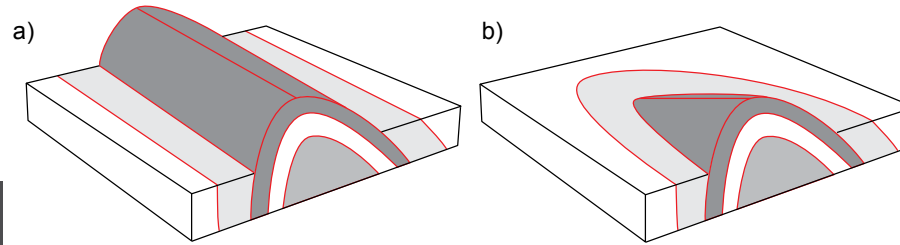


Figure 39: a) A non-plunging fold. b) A plunging fold.

Folds can be classified in a variety of ways. Here are just a few (for more detailed fold classifications see Fossen, 2010 chapter 11).

Orientation of a fold

Orientation of a fold is defined by the orientation of its axial plane (figure 38) and its hinge line, whether horizontal or plunging (figure 39). An **upright** fold has a vertical axial plane. An **inclined** fold has a dipping plane and a **recumbent** fold has a horizontal axial plane. A **non-plunging** fold has a horizontal hinge line, whilst a **plunging** fold has a inclined hinge line.

Where a hinge line is straight the fold is said to be **cylindrical**. Where a hinge line curves the fold is **non-cylindrical**. Curvature of the hinge line results in folds that close in the direction in which their limbs converge. Double plunging folds (or

pericline) form elongate domes or basins.

Tightness

The tightness of a fold is a measure of the interlimb angle, the angle between the limbs (figure 40) and relates to the amount of strain during deformation.

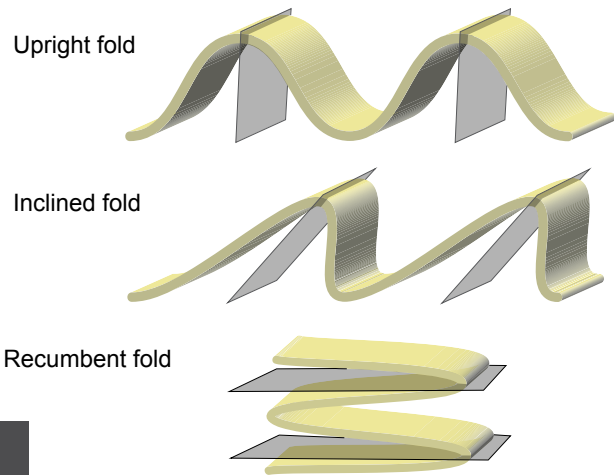


Figure 38: Orientation of fold axial plane.

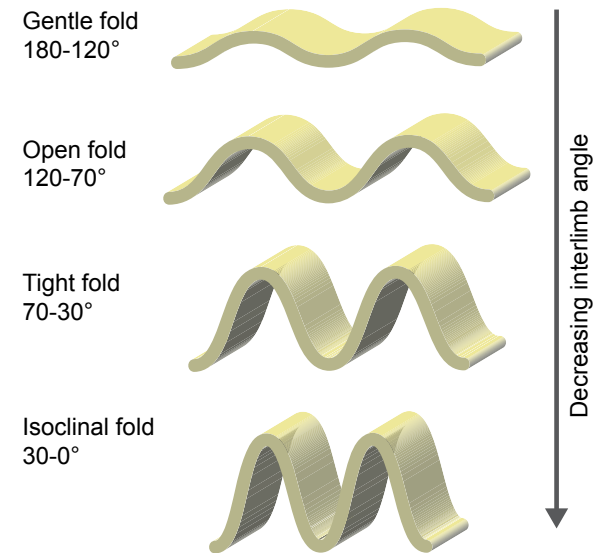


Figure 40: Interlimb angle: Folds become tighter with increased compression.



Figure 41: Open, upright syncline in Carboniferous limestone, Stackpole Quay, Pembrokeshire (E.Condliffe)



Figure 42: Tight, inclined folds, SW England (D.Paton)



Figure 43: Isoclinal recumbent fold in Helminthoid Flysch, Apennines (RDR)

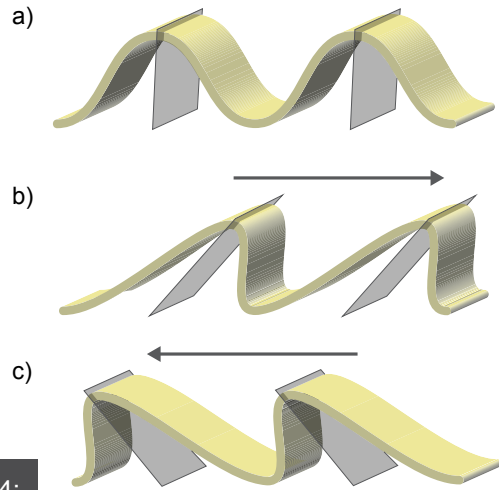


Figure 44:
a) Symmetric fold with neutral vergence. b) Asymmetric fold verging to the right. c) Asymmetric fold verging to the left.

Fold symmetry and vergence

The symmetry of a fold relates to its limb lengths (figure 44). **Symmetric** folds have equal limb lengths and the two sides of the fold are mirror images (figure 46). **Asymmetric** folds have a shorter and a longer limb (figures 47 and 48). A series of folds with the same asymmetry are said to have **vergence**. The direction of vergence is determined by the sense of displacement of the upper limb relative to the lower limb. When viewed down plunge, a fold verges to the right where there is apparent clockwise rotation of the short limb and to the left where there is apparent anti-clockwise rotation of the short limb.

Dip isogons

Dip isogons are a means of classifying folds based on the geometric relationship between their outer and inner profiles. Lines are drawn on the profile plane and connect points of equal dip between the inner and outer surfaces of the fold. These show the changes in layer thickness around a fold (figure 45).

Dip isogons reflect how a rock has responded to compression and can vary between different rock layers within the same fold. Dip isogons are divided into three groups on whether the isogons converge, diverge or are parallel as they cross from the outer to the inner arc of the fold.

Class 1: Dip isogons converge towards the inner arc. The inner arc is tighter than outer arc. Class 1 is divided into three sub-classes: Class 1a) Hinge zones are thinner than the limbs.

Class 1b) Parallel or concentric folds with constant layer thickness (figure 46). Class 1c) Slightly thickened hinge zones and narrower limbs. Class 2: Dip isogons are parallel to axial trace, known as similar or shear folds (figure 47). Class 3: Dip isogons diverge towards inner arc. The inner arc is more open than outer. This produces thinner limbs and thicker hinges.

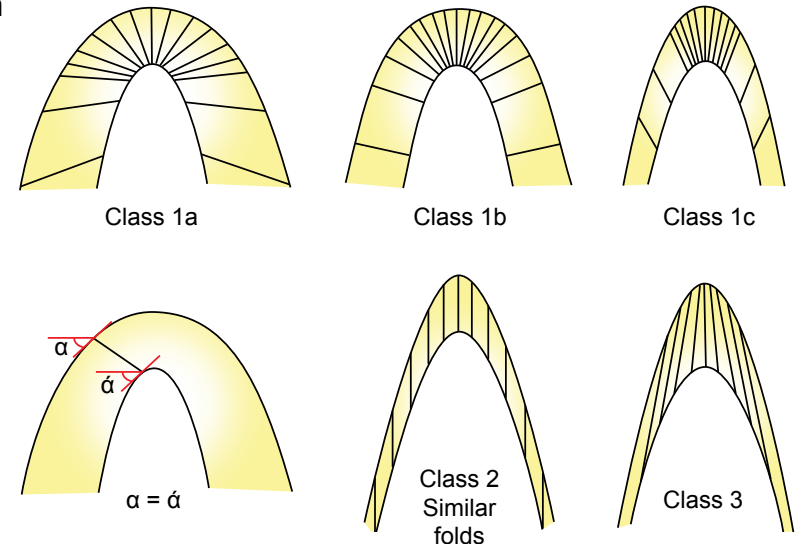


Figure 45: Different classes of dip isogons (after Ramsey, 1967).



Figure 46: Class 1b, minor folds with neutral vergence in the hinge zone of the Rhoscolyn anticline, Anglesey (J.Houghton).

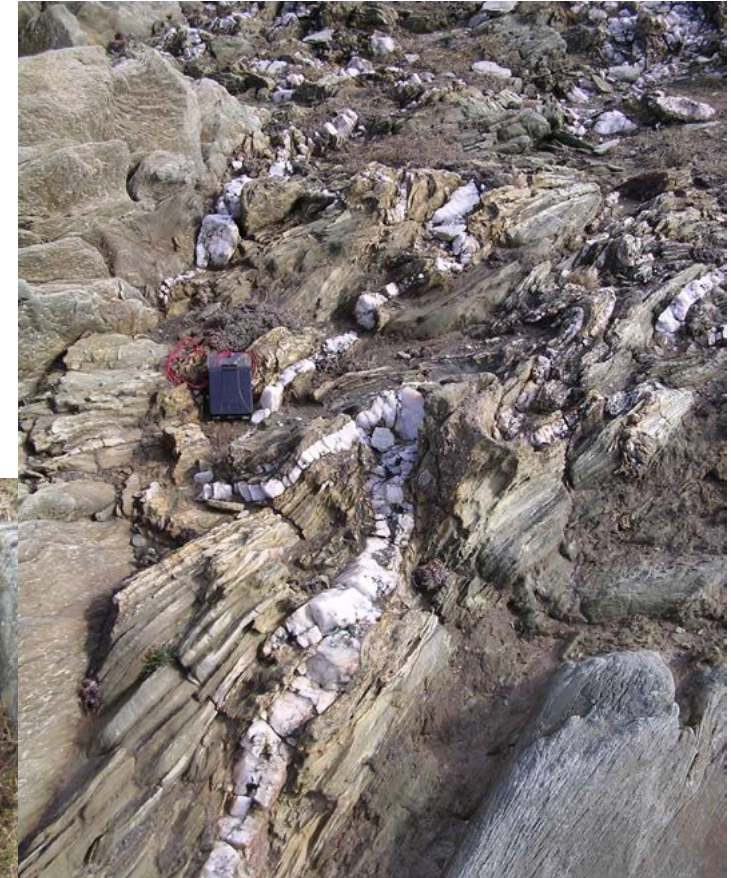


Figure 48: Left-verging minor folds in quartzite vein, Rhoscolyn, Anglesey (J.Houghton).



Figure 47: Right-verging, class 2 minor folds in marble (B.Yardley).

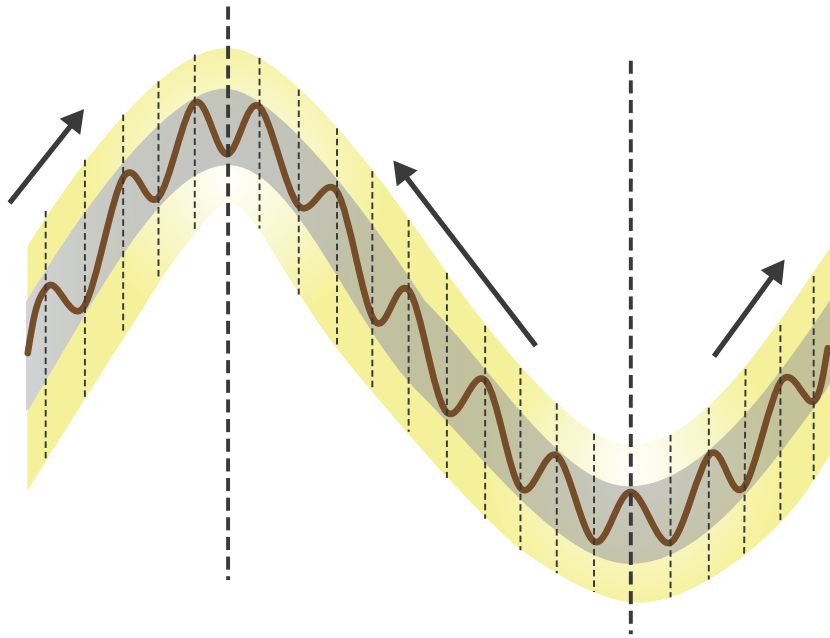


Figure 49: Geometric relationships between first and second order folds. Black arrows give direction of vergence of the minor folds.

Minor folds

Large scale (or first order) folds often have minor (second order) folds associated with them, which formed during the same phase of deformation. These folds form in the same stress field and so have the same geometric features. The axial planes of the minor and large scale folds are parallel (figure 49).

Within the hinge zones of the larger folds the

minor folds are symmetric. On the limbs of the large scale folds the minor folds are asymmetric. The minor folds verge towards the hinge zone of the larger scale antiform.

These geometric relationships are particularly useful in the field where the vergence of the minor folds can be used to predict the position of larger scale folds.

Kink bands and chevron folds

Kink bands develop in well layer anisotropic rocks. They form as asymmetric folds where segments of the layering rotate during deformation. The rotated segments have sharp boundaries and angular hinges. They often form in conjugate pairs with the opposite sense of rotation (figures 50 and 51).

Chevron folds likewise form in well layered rocks. They form by flexural slip between the layers.

They have sharp hinges and straight limbs (figures 50 and 52). Space problems as the layers fold lead to ductile flow of less competent layers into hinge zone (figure 53).

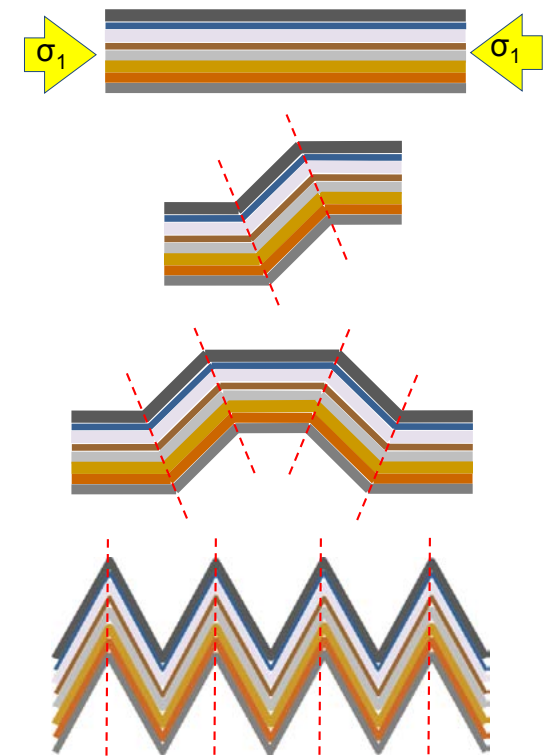


Figure 50: a) Layered rocks. b) Kink band. c) Conjugate pair of kink bands. d) Chevron folds



Figure 51: Kink band in siltstone, Whitesands Bay, Pembroke (J.Houghton).



Figure 53: Less competent shales flowing into the hinge zone of the chevron folds, SW England (D.Paton).



Figure 52: Large scale chevron folds, SW England (D.Paton).

How rocks fold

Rock layers accommodate layer parallel shortening by folding. This shortening may be taken up either along the limbs with slip concentrated between or within the layers (**flexural slip or flow**) or concentrated around the hinge zone (**tangential longitudinal strain**). In both cases bed thickness will be maintained.

Flexural slip and flow

In flexural slip and flexural flow strain is concentrated in limbs and dies out towards hinge (figure 54a and b). The limbs show opposite senses of shear. Flexural slip occurs in well layered rocks usually in the brittle regime. The slip is concentrated between beds or along incompetent layers (e.g. shale).

Flexural flow occurs in the ductile regime and strain is evenly distributed across the limb.

Tangential Longitudinal Strain (TLS)

Tangential longitudinal strain (also known as orthogonal flexure) occurs in more homogeneous,

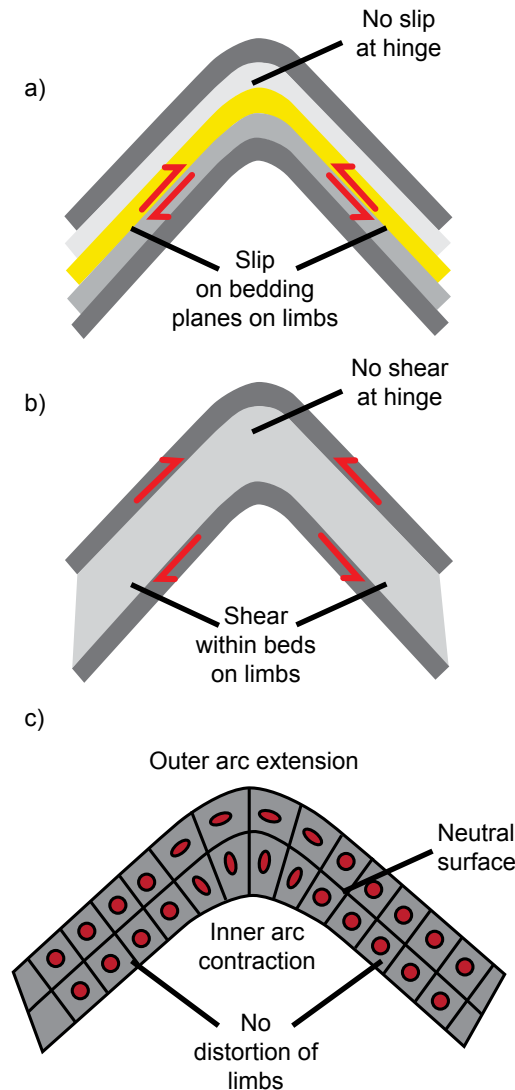


Figure 54: a) Flexural slip. b) Flexural flow. c) Tangential longitudinal strain

thicker, less well layers rocks (figure 54c). Strain is concentrated in hinge zone with no strain along limbs. It results in outer arc extension and inner arc contraction across the hinge zone. Areas of extension and contraction are separated by the neutral surface along which there is no strain. Veins often develop around the outer arc of the hinge zone where it is stretched and pressure solution cleavage (figure 55) or small scale thrust faults in the inner hinge zone where it is compressed.



Figure 55: Outer arc extension with wedge-shaped veins around the outside of the hinge zone (J.Houghton).

Cleavage

Cleavage is a secondary layering that develops in the rock as a result of deformation (figure 56). It forms perpendicular to sigma one, is often associated with folding and can be localized or regional in extent.

Cleavage is a foliation along which a rock can split. It occurs at the very lowest grades of metamorphism and as the grade of metamorphism increases it is replaced by schistosity

Cleavage develops by the rotation of platy minerals, growth of minerals in a preferred orientation and by pressure solution. The lithology of the rock is important as without phyllosilicates there is little to no cleavage development.

Pressure solution

Pressure solution occurs where soluble minerals go into solution in areas of high pressure and are re-precipitated in areas of lower pressure. Pressure solution cleavage forms where soluble minerals dissolve along grain boundaries perpendicular to

sigma one and are re-precipitated perpendicular to sigma three. During this process clay minerals are concentrated into bands. This results in a cleavage with fine layers of soluble and insoluble minerals giving the rock a banded appearance.

Slaty cleavage forms by both the rotation of minerals and pressure solution (figure 57). **Pressure solution cleavage** forms by pressure solution alone and is particularly common in limestones (figure 58).

Cleavage refraction

Cleavage refraction is the change in the angle of dip of cleavage between layers of different competence (figures 56c and 59). The cleavage will have a lower dip in the least competent beds and a higher dip in the more competent. Incompetent beds contain more phyllosilicates so tend to have better developed cleavage. The cleavage forms perpendicular to sigma one but then is rotated by localized shearing along the bedding plane during deformation.

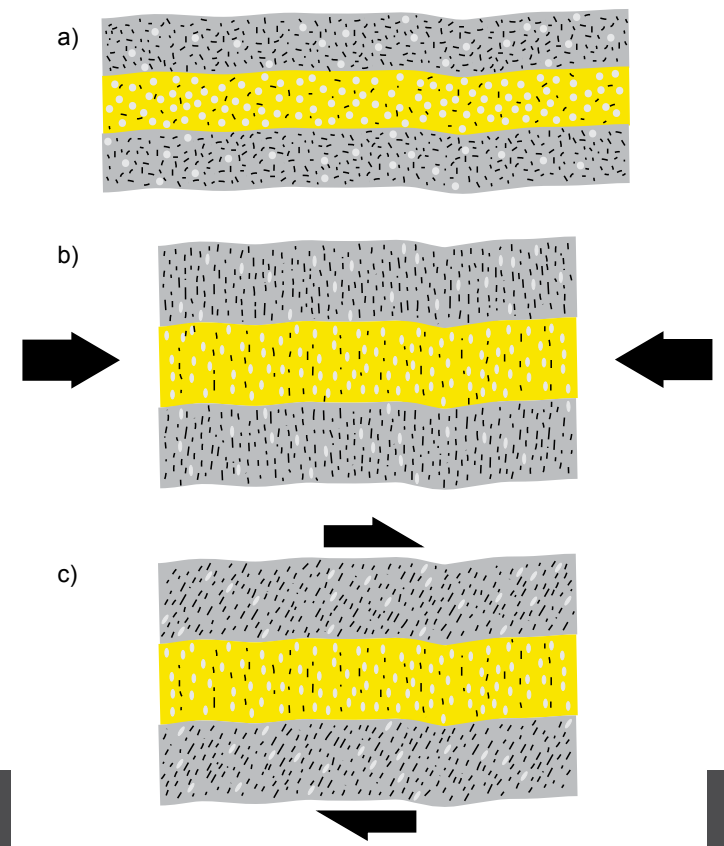


Figure 56: a) Before deformation. b) After deformation. The grey layers are phyllosilicate-rich and develop a strong cleavage. The yellow layer has fewer platy minerals and a weaker cleavage develops. c) Cleavage refraction.



Figure 57: Slaty cleavage Note how the cleavage has a lower dip than the bedding. This is because the beds here are on the overturned limb of a fold, Abereiddi Bay, Pembrokeshire (G.McLeod).



Figure 58: Pressure solution cleavage showing dark insoluble bands and light soluble bands (R. Butler).



Figure 59: Cleavage refraction. Coarser-grained pale layers have less well developed, steeply dipping cleavage. Finer-grained dark grey layers have better developed, shallower dipping cleavage. En echelon veins in the darker layer suggests this layer has undergone shearing (E.Condliffe).

Relationship between folding and cleavage

Fold axial planes and cleavage both form perpendicular to the direction of maximum compression (σ_1). Where folding and cleavage form in the same stress field the cleavage planes and fold axial planes will be approximately parallel to each other (figure 60). The orientation of the cleavage will be constant across the fold but, because the dips of the beds vary, the angle between bedding and cleavage will change in a consistent manner round the fold.

can be seen this relationship can be used to establish the geometry of the fold. In the same way that axial planes of minor folds verge towards the hinge zone of the major antiform, cleavage on fold limbs verges towards the antiform hinge zone. Look at the angle between bedding and cleavage in figure 60. In hinge zones the cleavage is perpendicular to bedding; it has neutral vergence. Along the limbs the cleavage is at an angle to bedding. Where it appears rotated clockwise relative to bedding it verges to the right and where it appears rotated anticlockwise it verges to the left.

At outcrop where only bedding and cleavage

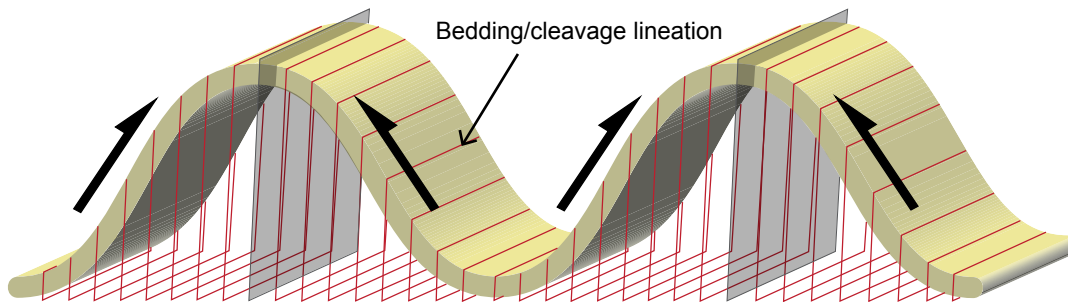


Figure 60: The relationship between cleavage and folding. Black arrows give sense of vergence. See text for details.

Where the cleavage plane cuts the bedding it forms a lineation on the bedding surface; the bedding cleavage intersection

lineation. Where the cleavage is associated with folding this lineation is parallel to the fold axis/hinge line.



Figure 61: Left-verging cleavage. Bedding cleavage intersection lineation is visible on the bedding plane (G.Lloyd).



Figure 62: Axial planar cleavage (M. Miller).



Figure 63: Axial planar cleavage, Anglesey (G. Lloyd).

Acknowledgments and references

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