University of Diyala
College of Science
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Lectures in Structural Geology
Prepared by:

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Lecture one

The terms **Structural geology** and **Tectonics** are derived from similar roots. **Structure** come from Latin word *struere*, which means to build, and **tectonics** from Greek word *tektos*, which means builder, the reference being to the motions and processes that build the crust of the Earth.

Structural Geology: Deals with the *origin*, *geometry* and *kinematics of structures' formation*. It requires an ability to visualize objects in three dimensions Fig. (1-1).

Fig.(1-1)

Plate Tectonics: Deal specifically with *plate generation*, *motion*, and *interaction* Fig.(1-2).

Tectonic structures: are produced in rocks in response to stress generated by plate motion within the Earth. They include all kinds of faults, folds along with other structures. They make up the tectonic framework of the earth.

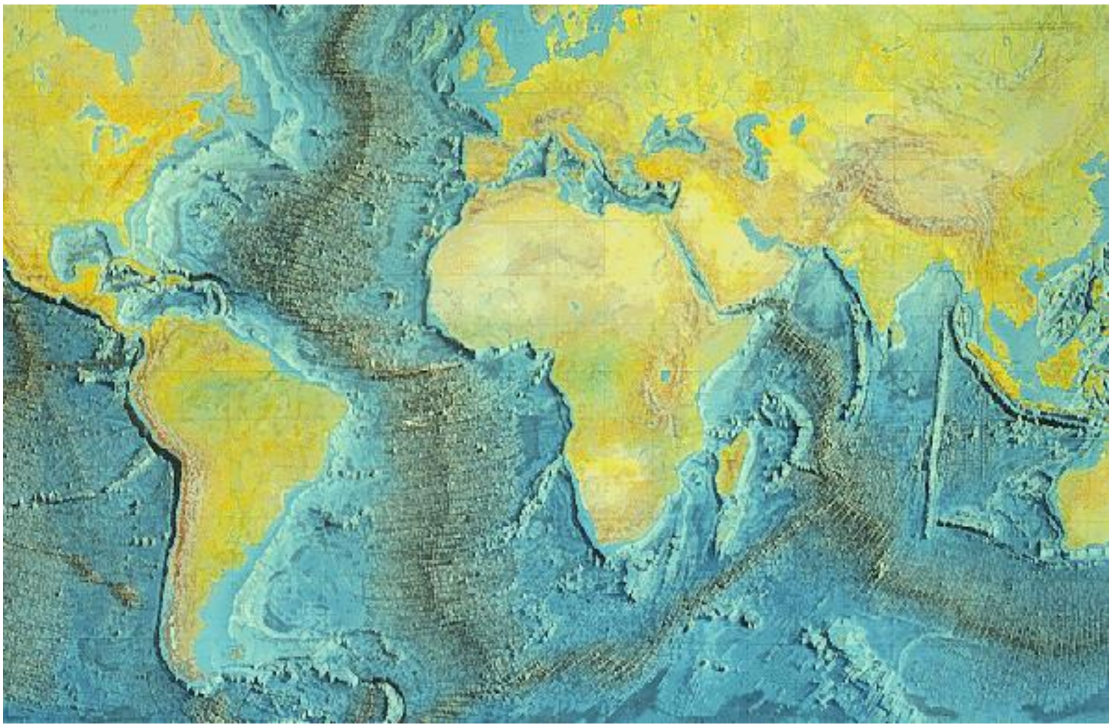


Fig.(1-2)

Structural Geology: Involve studying outcrop and microscopic size local structures such as: *Non tectonic structures* and *Folds, faults,* and other related structure Fig. (1-3).

Fig. (1-3)

Tectonics: involve the study of larger features, and regional structures such as: *mountain ranges, parts of continents, trenches, island arcs, and oceanic ridges* Fig(1-4)**.**

Fig.(1-4)

1.1 ROCK DEFORMATION

STRESS

Stress is a force exerted against an object. Tectonic forces exert different types of stress on rocks in different geologic environments. The first, called **confining stress** or **confining pressure**, occurs when rock or sediment is buried (Fig.1-5a).Confining pressure merely compresses rocks but does not distort them, because the compressive force acts equally in all directions, like water pressure on a fish. Burial pressure compacts sediment and is one step in the lithification of sedimentary rocks. Confining pressure also contributes to metamorphism during deep burial in sedimentary basins. In contrast, **directed stress** acts most strongly in one direction. Tectonic processes create three types of directed stress. **Compressive stress** squeezes rocks together in one direction. It frequently acts *horizontally, shortening the distance parallel to the squeezing direction* (Fig.1-5b). **Compressive stress** is common in convergent plate boundaries, where two plates converge and the rock crumples, just as car fenders crumple during a head-on collision. **Extensional stress** (often called **tensional stress**) pulls rock apart and is the opposite of tectonic compression (Fig.1-5c). Rocks at a divergent plate boundary stretch and pull apart because they are subject to extensional stress. **Shear stress** acts in parallel but opposite directions (Fig.1-5d). Shearing deforms rock by causing one part of a rock mass to slide past the other part, as in a transform fault or a transform plate boundary.

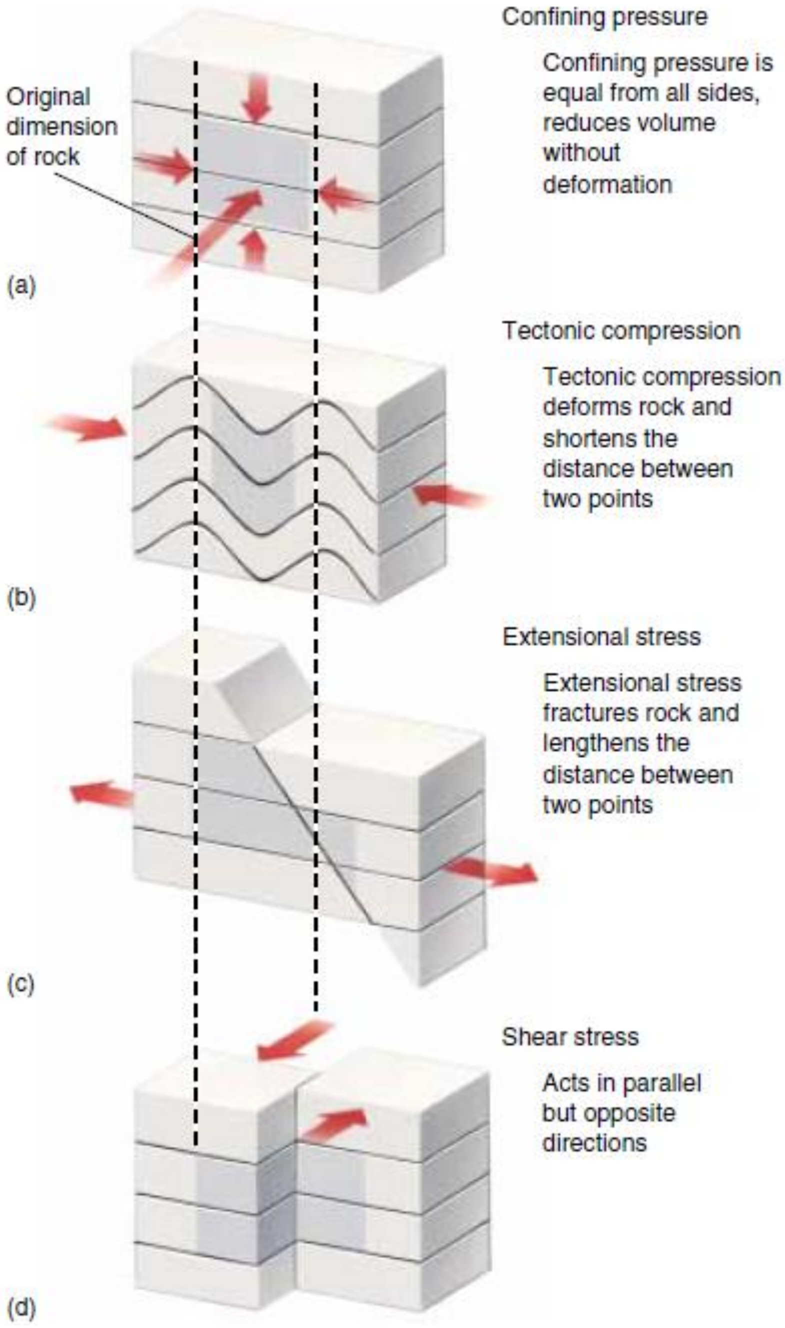


Figure (1-5) (a) Confining pressure acts equally on all sides of a rock. Thus, the rock is compressed much as a balloon is compressed if held under water. Rock volume decreases without deformation. (b) Tectonic compression shortens the distance parallel to the stress direction. Rocks fold or fracture to accommodate the shortening. (c) Extensional stress lengthens the distance parallel to the stress direction. Rocks commonly fracture to accommodate the stretching. (d) Shear stress deforms the rock parallel to the stress direction.

STRAIN

Strain is the deformation produced by stress. A rock responds to tectonic stress by *elastic deformation*, *plastic deformation*, or *brittle fracture*. An elastically deformed rock springs back to its original size and shape when the stress is removed. During plastic deformation, a rock deforms like putty and retains its new shape. In some cases a rock will deform plastically and then fracture (Fig. 1-6).

Figure 1-6 This rock (in the Nahanni River, Northwest Territories, Canada) folded plastically and then fractured.

Factors That Control Rock Behavior

Several factors control whether a rock responds to stress by elastic or plastic deformation or fails by brittle fracture:

1. ***The nature of the material***. Think of a quartz crystal, a gold nugget, and a rubber ball. If you strike quartz with a hammer, it shatters. That is, it fails by brittle fracture. In contrast, if you strike the gold nugget, it deforms in a plastic manner; it flattens and stays flat. If you hit the rubber ball, it deforms elastically and rebounds immediately, sending the hammer flying back at you. Initially, all rocks react to stress by deforming elastically. Near the Earth's surface, where temperature and pressure are **low**, different types of rocks behave differently with continuing stress. Granite and quartzite tend to behave in a brittle manner. Other rocks, such as shale, limestone, and marble, have greater tendencies to deform plastically.

2. ***Temperature***. The higher the temperature, the greater the tendency of a rock to behave in a plastic manner. It is difficult to bend an iron bar at room temperature, but if the bar is heated in a forge, it becomes plastic and bends easily.

3. *Pressure*. High confining pressure also favors plastic behavior. During burial, both temperature and pressure increase. Both factors promote plastic deformation, so deeply buried rocks have a greater tendency to bend and flow than shallow rocks.

4. *Time*. Stress applied over a long time, rather than suddenly, also favors plastic behavior. Marble park benches in New York City have sagged plastically under their own weight within 100 years. In contrast, rapidly applied stress, such as the blow of a hammer, to a marble bench causes brittle fracture.

GEOLOGIC STRUCTURES

Enormous compressive forces can develop at a convergent plate boundary, bending and fracturing rocks in the tectonically active region. In some cases the forces deform rocks tens or even hundreds of kilometers from the plate boundary. Because the same tectonic processes create great mountain chains, rocks in mountainous regions are commonly broken and bent. Tectonic forces also deform rocks at divergent and transform plate boundaries.

A **geologic structure** is any feature produced by rock deformation. Tectonic forces create three types of geologic structures: **folds**, **faults**, and **joints**.

FOLDS

A **fold** is a bend in rock (Fig. 1-7). Some folded rocks display little or no fracturing, indicating that the rocks deformed in a plastic manner.

Figure 1-7 A fold is a bend in rock. These are in quartzite in the Maria Mountains, California.

In other cases, folding occurs by a combination of plastic deformation and brittle fracture. Folds formed in this manner exhibit many tiny fractures. If you hold a sheet of clay between your hands and exert compressive stress, the clay deforms into a sequence of folds (Fig. 1-8). This demonstration illustrates three characteristics of folds:

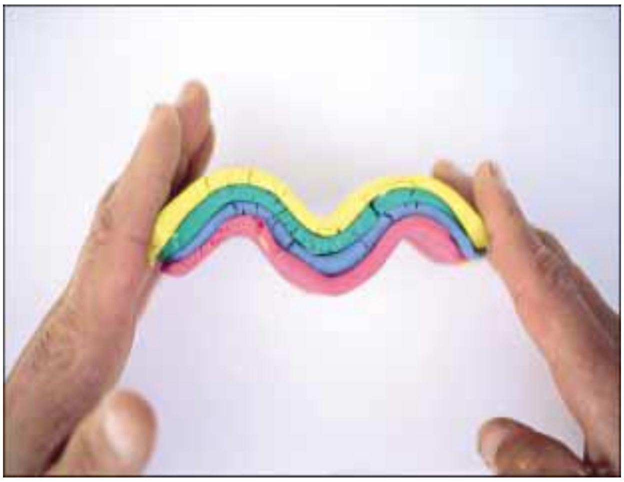


Figure 1-8 Clay deforms into a sequence of folds when compressed

1. Folding usually results from *compressive stress*. For example, tightly folded rocks in the Himalayas indicate that the region was subjected to compressive stress.

2. Folding always *shortens the horizontal distances* in rock. Notice in Figure 1-9 that the distance between two points, A and A′, is shorter in the folded rock than it was before folding.

3. Folds usually occur as a *repeating pattern* of many folds as in the illustration using clay.

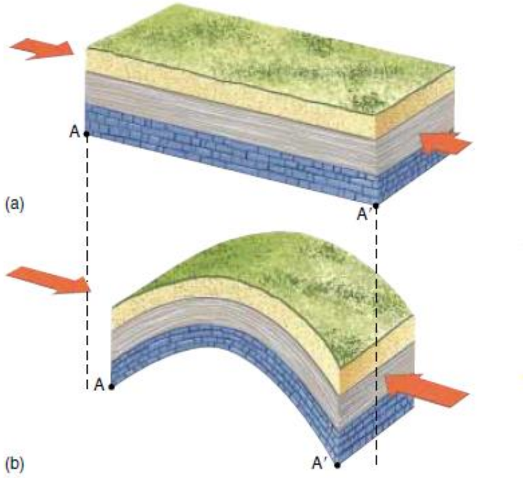


Figure 1-9 (a) Horizontally layered sedimentary rocks b) A fold in the same rocks. The forces that folded the rocks are shown by the arrows. Notice that points A and A′ are closer after folding.

Simple folds are divided into two types, that is, **anticlines** and **synclines** in the former, the beds are convex upwards, whereas in the latter, they are concave upwards. In the anticline when we move toward the core we can show the oldest rocks in contrary to the syncline we show the youngest rocks. The **crestal line** of an anticline is the line that joins the highest parts of the fold, whereas the **trough line** runs through the lowest parts of a syncline (Fig. 1.10a, b).

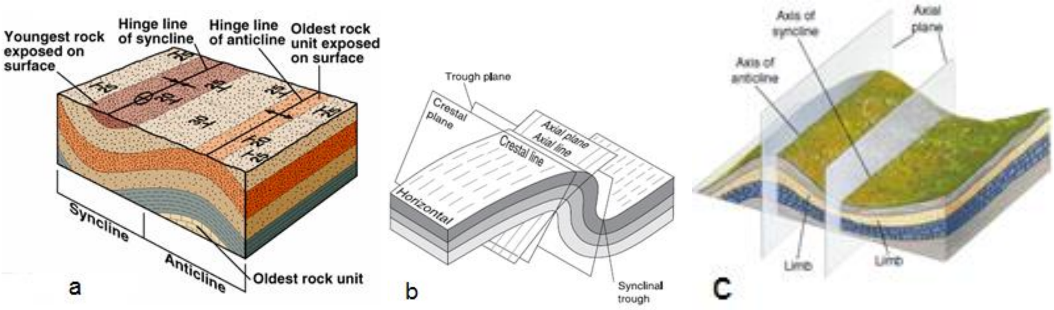


Figure 1-10 (a) symmetrical anticline and syncline, and the parts of a fold. (b) Asymmetrical anticline and syncline. (c) Axial plane of anticline and axial plane of syncline .

The **hinge line** of a fold is the line along which the greatest curvature exists and can be either straight or curved. However, the **axial line** is another term that has been used to describe the hinge line. The **limb** of a fold occurs between the hinges, all folds having two limbs. The **axial plane** of a fold is commonly regarded as the plane that bisects the fold and passes through the hinge line (Fig. 1.10c).
A fold arching upward is called an **anticline** and one arching downward is a **syncline**. The sides of a fold are called the **limbs**. Notice that a single limb is shared by an anticline–syncline pair. A line dividing the two limbs of a fold and running along the crest of an anticline or the trough of a syncline is the **fold axis**. The **axial plane** is an imaginary plane that runs through the axis and divides a fold as symmetrically as possible into two halves. In many folds, the axis is horizontal, as shown in Figure 1–10a, b. If you were to walk along the axis of a horizontal anticline, you would be walking on a level ridge.

In other folds, the axis is inclined or tipped at an angle called the **plunge**, as shown in Figure 1–11. A fold with a plunging axis is called a **plunging fold**. If you were to walk along the axis of a plunging fold, you would be traveling uphill or downhill along the axis. Even though an anticline is structurally a high point in a fold, anticlines do not always form topographic ridges. Conversely, synclines do not always form valleys Figure 1–11B .

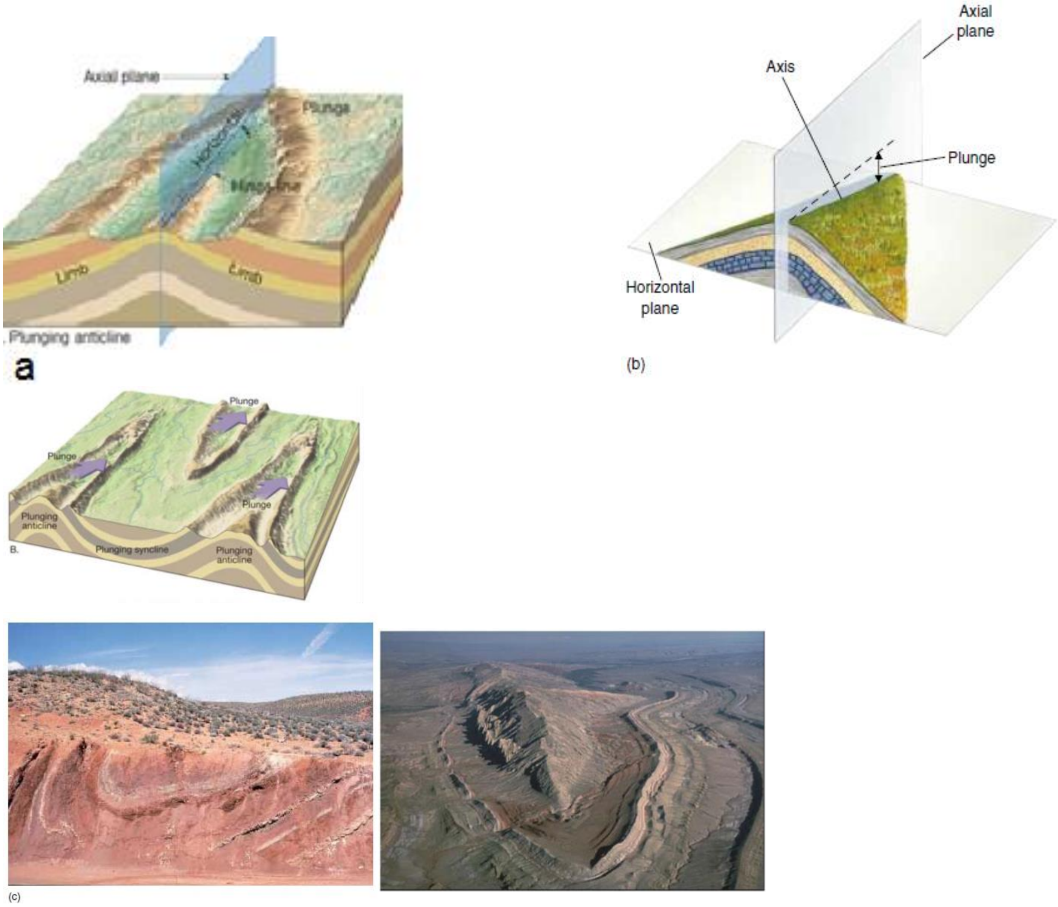
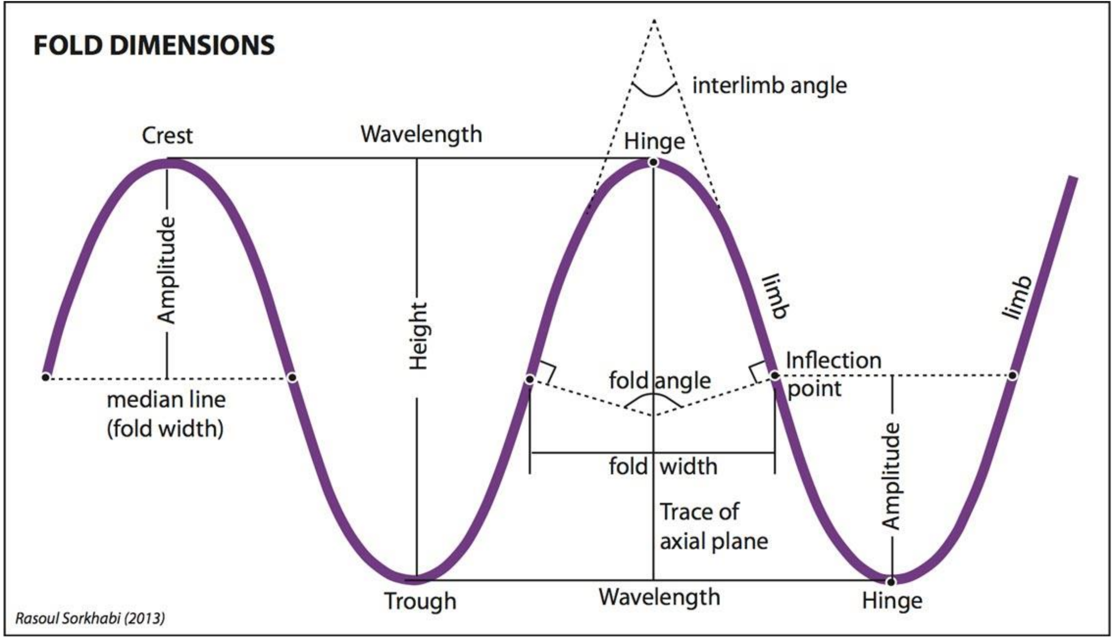


Figure 1-11 Plunging anticline and plunging syncline, figures and photos.

Landforms are created by combinations of tectonic and surface processes. The **amplitude** of a fold is defined as the vertical difference between the crest or the trough and the median line, whereas the **wave length of a fold** is the horizontal distance from crest to crest or trough to trough.

Fig 1-11 B, shows a syncline lies beneath the mountain peak and an anticline forms the low point, or saddle, in the Canadian Rockies, Alberta.



Types of Folding

Corresponding to the dip angle of the limb and the axial plane the folds are classified to four types, the first is **symmetrical** if both limbs are arranged equally about the axial plane so that the dips on opposing flanks are the same; otherwise they are **asymmetrical** (Fig. 1.12). In symmetrical folds, the axial plane is vertical, whereas it is inclined in asymmetrical folds. As folding movements become intensified, **overturned folds** are formed in which both limbs are inclined, together with the axial plane, in the same direction but at different angles (Fig. 1.12). In a **recumbent fold**, the beds have been completely overturned so that one limb is inverted, and the limbs, together with the axial plane, dip at a low angle to the same direction (Fig. 1.12).

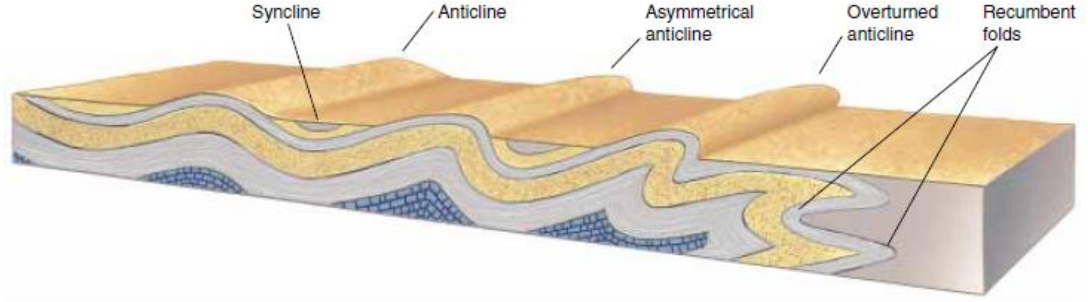


Figure 1-12 Cross-sectional view of five different kinds of folds. Folds can be symmetrical, as shown on the left, or asymmetrical, as shown in the center. If a fold has tilted beyond the perpendicular, it is overturned.

According to the **shape** of the fold, many types have been recognized such as:

Parallel or concentric folds are those where the strata have been bent into more or less parallel curves in which the **thickness** of the individual beds remains the same. From Figure 1.13a, it can be observed that, because the thickness of the beds remains the same on folding, the shape of the folds changes with depth and, in fact, they fade out. Parallel folding occurs in competent (relatively strong) beds that may be interbedded with incompetent (relatively weak, plastic) strata, Fig. 1.13a and 1.13A.
Similar folds are those that retain their shape with depth. This is accomplished by flowage of material from the limbs into the crest and trough regions (Fig. 1.13b). Similar folds are developed in incompetent strata. However, true similar folds are rare in nature, for most change their shape to some degree along the axial plane. Most folds exhibit both the characteristics of parallel and similar folding Fig. 1.13 B.

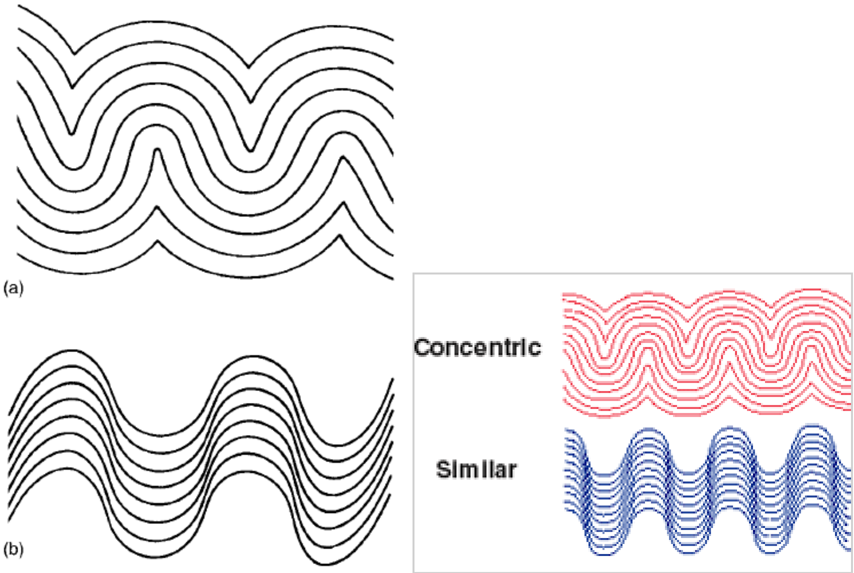


Figure 1.13 Shows (a) Parallel folding. (b) Similar folding.

Fig. 1.13 A. These are folds in Cretaceous strata exposed at Ernst Tinaja in Big Bend National Park. They show the disharmonic geometry and maintenance of bedding-perpendicular thickness that characterize a parallel fold style.

Fig. 1.13B. Shows photo of similar fold.

Several other common terms specify relative orientations of the limbs of the folds. A **homocline** comprises a surface, such as bedding, that has a uniform nonhorizontal attitude over a regional scale with no major fold hinge (Fig.1.14A). A **monocline** is a special type of fold with only one limb or a fold pair that has two long horizontal limbs connected by a relatively short limb (Fig.1.14B). A monocline may develop where sedimentary rocks sag over an underlying fault (Fig.1.15). A **structural terrace** is a fold pair with two long planar inclined limbs connected by relatively short horizontal limb (Fig.1.14C). A **recumbent** fold in which one limb is overturned i.e. rotated more than 90° (Fig.1.14D).

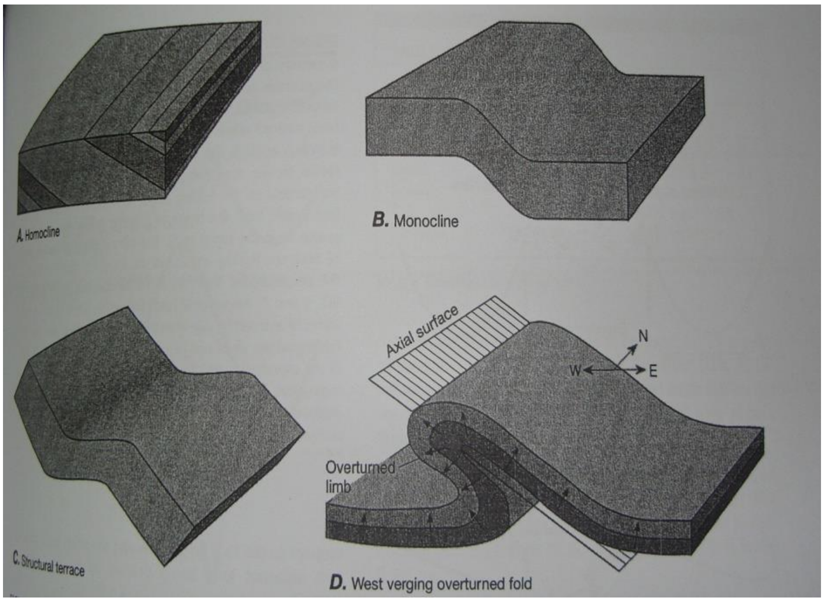


Fig. 1.14 Structural terms describing the orientation of fold limds.

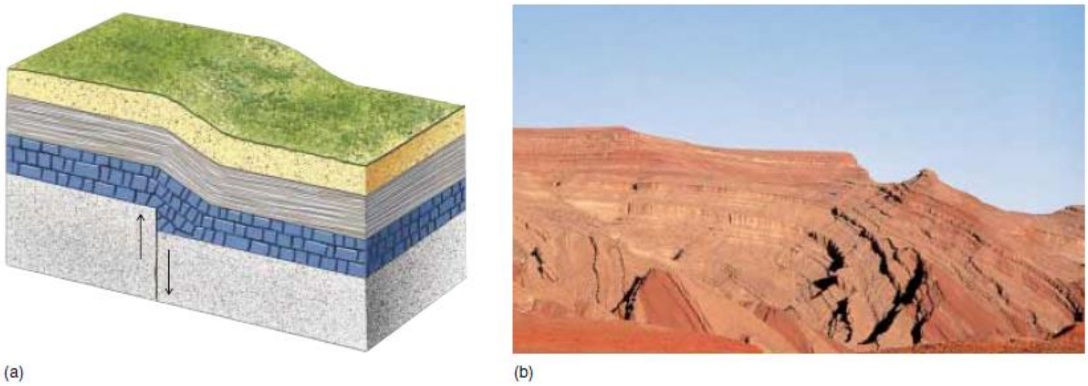


Figure 1–15 (a) A monocline formed where near-surface sedimentary rocks sag over a fault. (b) A monocline in southern Utah.

Zigzag or chevron folds have straight or nearly straight limbs with sharply curved or even pointed hinges (Fig. 1.16A and 1.17). Such folds possess features that are characteristic of both parallel and similar folds in that the strata in their limbs remain parallel, beds may be thinned but they never are thickened, and the pattern of the folding persists with depth. **Box fold** is one in which the crest broad and flat; two hinges are present, one on either side of the flat crest (Fig. 15B).

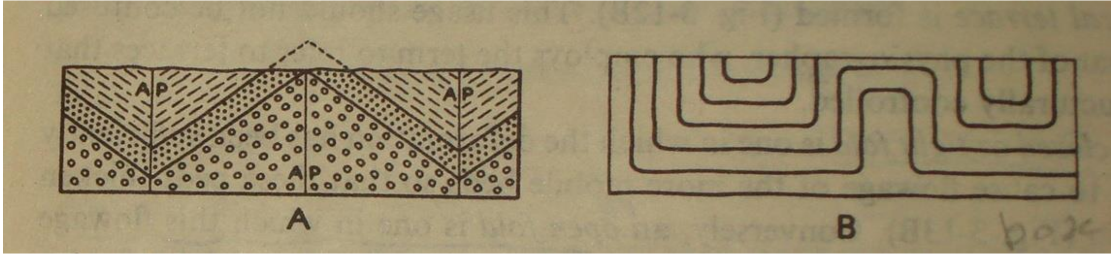


Fig. 16. Shows some varieties of folds. AP, axial plane. (A) Chevron fold. (B) Box fold.

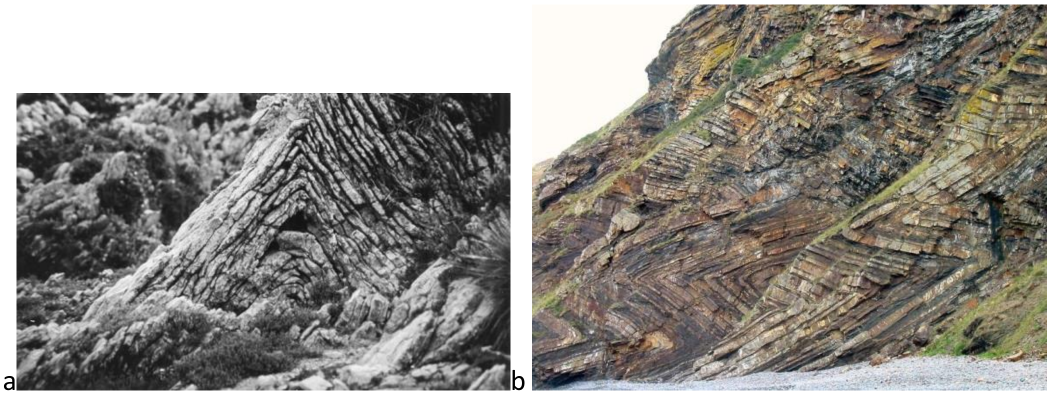


Fig.1.17(a) Chevron fold in limestone of Miocene age, Kaikuora, South Island, New Zealand, **b**, Chevron folds with flat-lying axial planes, Millook Haven, North Cornwall, UK

Fan fold is one in which both limbs are overturned (Fig.1.18A). In the anticlinal fan fold, the two limbs dip toward each other; in the synclinal fan fold, the two limb dip away from each other.

Kink bands are narrow bands, usually only a few inches or few feet wide, in which the beds assume a dip that is steeper or gentler than that in the adjacent beds (Fig.18B).

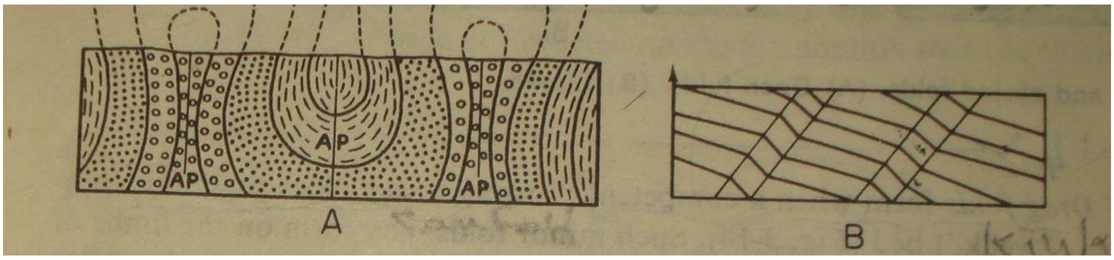
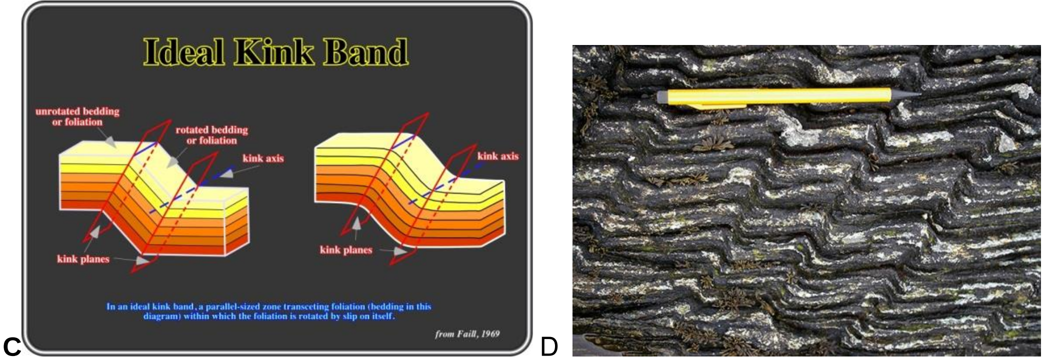


Fig.18. Shows some varieties of folds. AP, axial plane. (A) Fan fold (B) Kink bands. A fracture may separate the kink band from the rest of beds.



In an ideal kink band, a parallel-sized zone transceting foliation (bedding in this diagram) within which the foliation is rotated by slip on itself.

Fig.18.C, modeling of kink band formation, D , photo of kink band figure.

Drag folds form when a competent (strong) bed slides past an incompetent (weak) bed, minor folds may form on the limbs of larger folds because of the slipping of beds past each other. The axial planes of the drag folds are not perpendicular to the bedding of the competent strata, but are inclined at an angle (Fig.1.19).

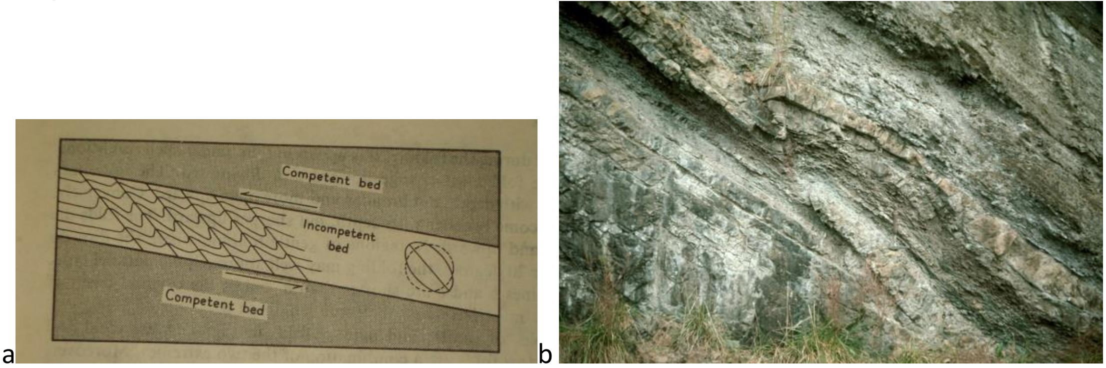
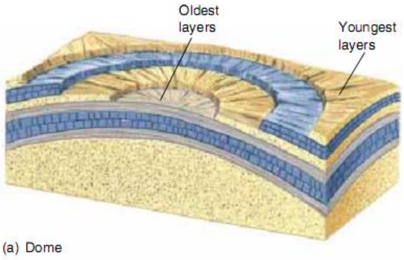


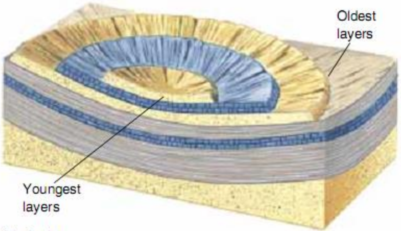
Fig.1.19. a, Mechanism of drag folds resulting from shearing, b, photo of drag folds in incompetent bed.

Dome

A circular or elliptical anticlinal structure resemble inverted bowls is called a **dome**. Sedimentary layering dips away from the center of a dome in all directions (Fig. 1.20a). A similarly shaped syncline is called a **basin (Fig. 1.20b)**. Domes and basins can be small structures only a few kilometers in diameter or less. Frequently, however, they are very large and are caused by broad upward or downward movement of the continental crust. The Black Hills of South Dakota are a large structural dome. The Michigan basin covers much of the state of Michigan, and the Williston basin covers much of eastern Montana, northeastern Wyoming, the western Dakotas, and southern Alberta and Saskatchewan.



(a) Dome

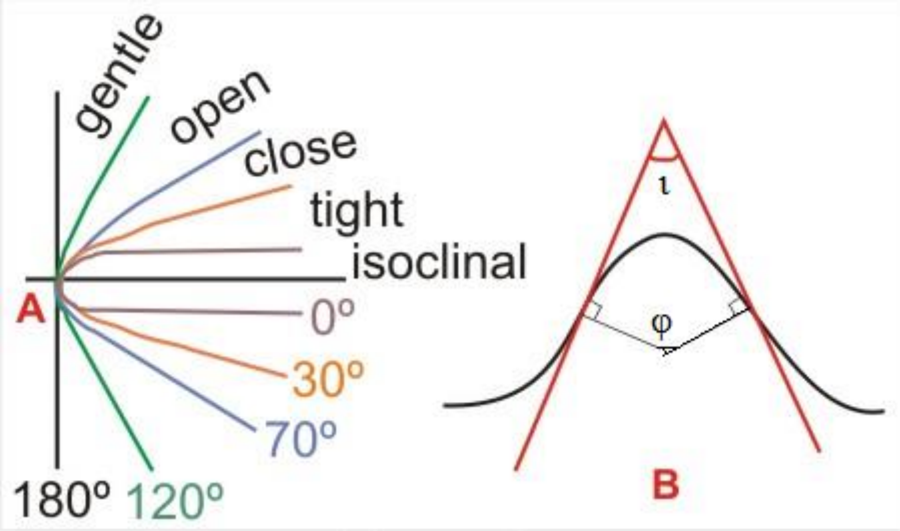


(b) Basin

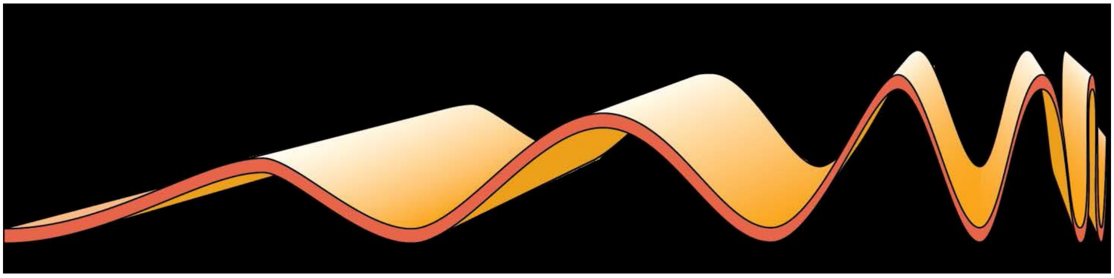
Figure 1.20 (a) Sedimentary layering dips away from a dome in all directions, and the outcrop pattern is circular or elliptical. (b) Layers dip toward the center of a basin.

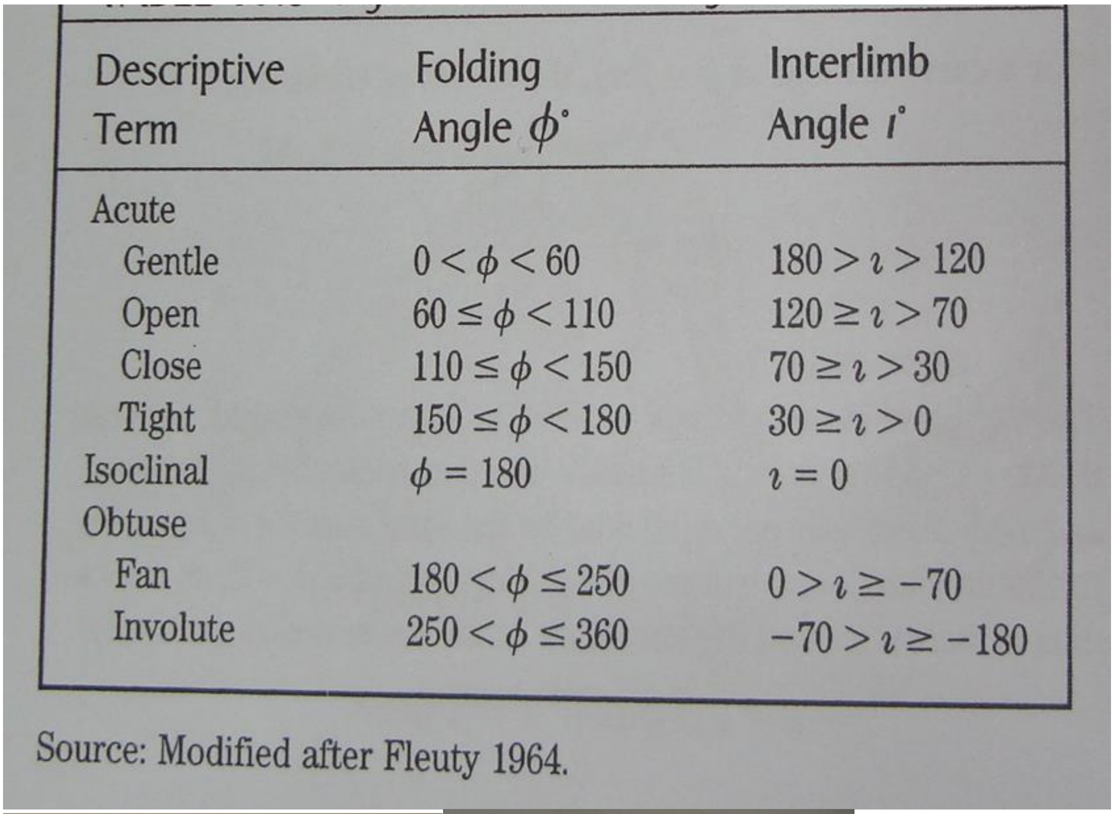
Fold Tightness

Fold tightness is defined by the interlimb angle (ι) or folding angle (φ). The **inter-limb angle**, which is the angle measured between the two projected planes from the limbs of the fold, it is the angle subtended by the tangents at two adjacent inflection points, which may reflect the intensity of compression. And the **folding angle** φ is the angle between the normals to the folded surface constructed at the inflection points.



They can be used to assess the degree of closure of a fold. Seven degrees of closure can be distinguished based on the inter-limb angle. **Gentle folds** are those with an inter-limb angle greater than 120˚; **open folds**, the inter-limb angle is between 120˚ and 70˚; **close folds**, it is between 70˚ and 30˚; **tight folds** are those with an inter-limb angle of less than 30˚; **isoclinal folds**, the limbs are parallel and so the inter-limb angle is zero; **fan folds** the inter-limb angle is between 0˚ and -70˚ and finally **Involute folds** the inter-limb angle is between -70˚and -180˚ ,table 1.





Descriptive Term	Folding Angle ϕ°	Interlimb Angle $\imath^\circ$
Acute		
Gentle	$0 < \phi < 60$	$180 > \imath > 120$
Open	$60 \leq \phi < 110$	$120 \geq \imath > 70$
Close	$110 \leq \phi < 150$	$70 \geq \imath > 30$
Tight	$150 \leq \phi < 180$	$30 \geq \imath > 0$
Isoclinal	$\phi = 180$	$\imath = 0$
Obtuse		
Fan	$180 < \phi \leq 250$	$0 > \imath \geq -70$
Involute	$250 < \phi \leq 360$	$-70 > \imath \geq -180$

Source: Modified after Fleuty 1964.

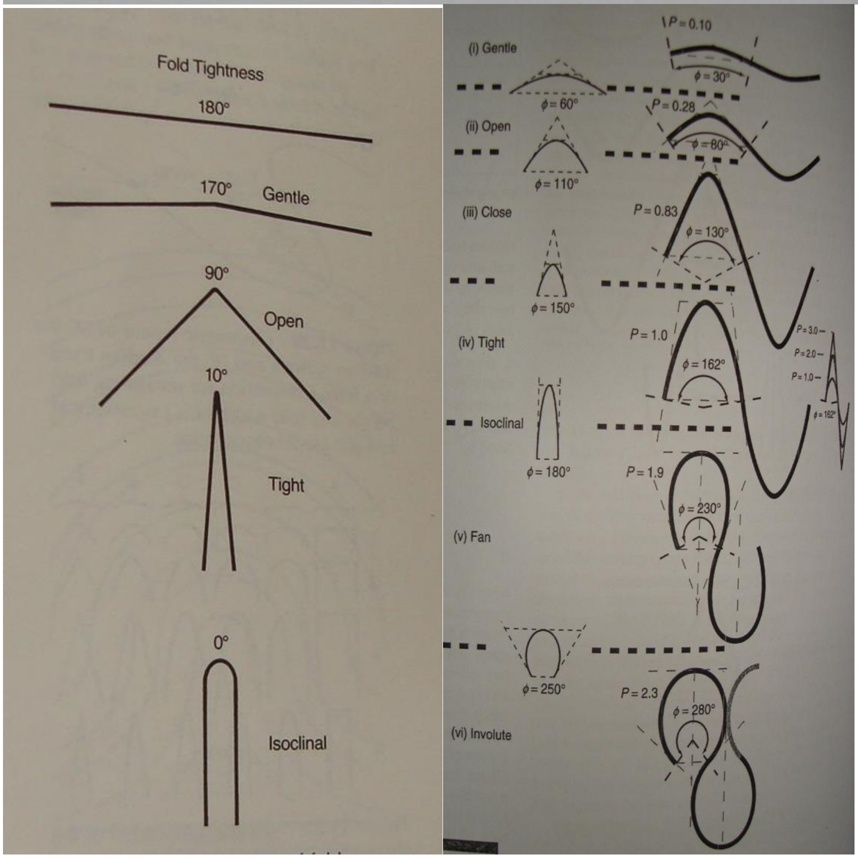


Fig.1.20. A. Old tightness classification. B. Modern tightness classification where **P=A/M**; is the ratio of the amplitude **A** of a fold measured along the axial surface, to the distance **M** measured between the adjacent inflection points that bound the fold.

Fold classification using Isogons

This method is based on the construction of dip isogons: line joining points of equal dip on either side of the folded layer. Using three geometric parameters are as follow (1) The dip isogons; (2) the orthogonal thickness $t\alpha$, which is the perpendicular distance between the two parallel tangents; (3) the axial trace thickness $T\alpha$, which is the distance between the two tangents measured parallel to the axial surface trace(Fig.1.21). The two measures of layer thickness $t\alpha$ and $T\alpha$ are related by

$$t\alpha = T\alpha \cos\alpha$$

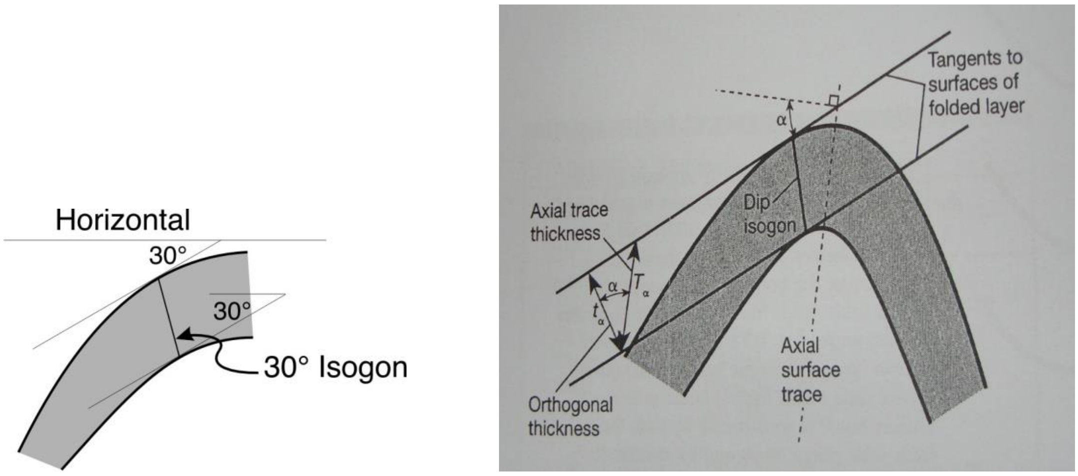


Fig.1.21 Definition of the layer inclination α, the dip isogons, the orthogonal thickness tα, the axial trace thickness Tα used to define the style of folded layer.

If the lines of dip isogons converge toward the inner side of the fold, that is convergent isogons; if they diverge toward the inner surface, that is the divergent isogons; and when they are parallel, that is parallel isogons.

Three classes of folds have been recognized (Fig.1.22):

Class 1, Convergent isogons imply that the **inner arc curvature** exceed that of the **outer arc**, which are subdivided in to three subclasses.

Sub-class 1A: strongly convergent

Sub-class 1B: parallel fold with isogons perpendicular to layering.

Sub-class 1C: weakly convergent.

Class 2, parallel isogons, and similar fold, the lines of isogons are parallel to the axial surface.

Class 3, Divergent isogons, imply that the outer arc curvature exceed that of the inner arc.

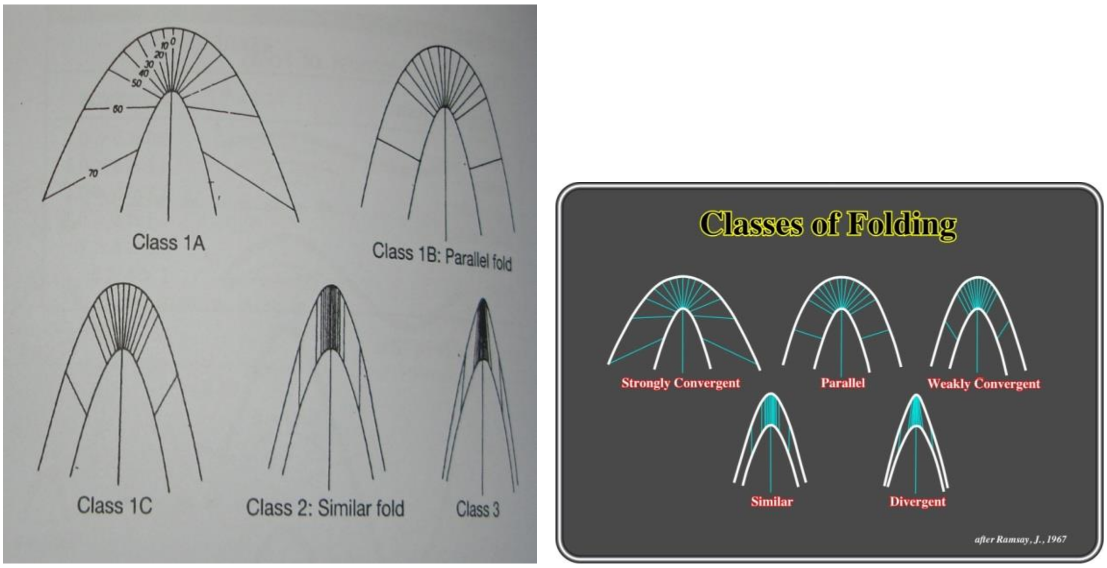


Fig.1.22. The dip isogon characteristics of the main fold classes.

For **class 1A folds**, the orthogonal thickness **tα** and the axial trace thickness **Tα** increase from hinge to limb, for **class1B folds**, the orthogonal thickness is constant from hinge to limb and the axial trace thickness **Tα** increases from hinge to limb, these folds also are commonly referred to as parallel folds because **tα** is constant all around fold, and for **class 1C folds**, the orthogonal thickness decrease from hinge to limb and the axial trace thickness **Tα** increase from hinge to limb.

In **class 2** and **class 3 folds**, the orthogonal thickness **tα** decrease from hinge to limb, but the axial trace thickness **Tα,** is constant in **class** 2 and decrease in **class 3 folds** from hinge to limb **Fig. 1.23.**

Classification of folds based on the orientation of hinge line and the axial surface

This type of folds classification depends on the combinations between the orientation of the hinge line and the axial surface orientation, as shown in table (2) and fig 1.24, when the axial surface is vertical, three types of hinge line orientations are formed as a **Horizontal normal**, **plunging normal** and **vertical**; when the axial surface is dipping, also three types are formed. They are **horizontal inclined**, **plunging inclined** and **reclined** and finally when the axial surface is horizontal, **recumbent fold** is formed.

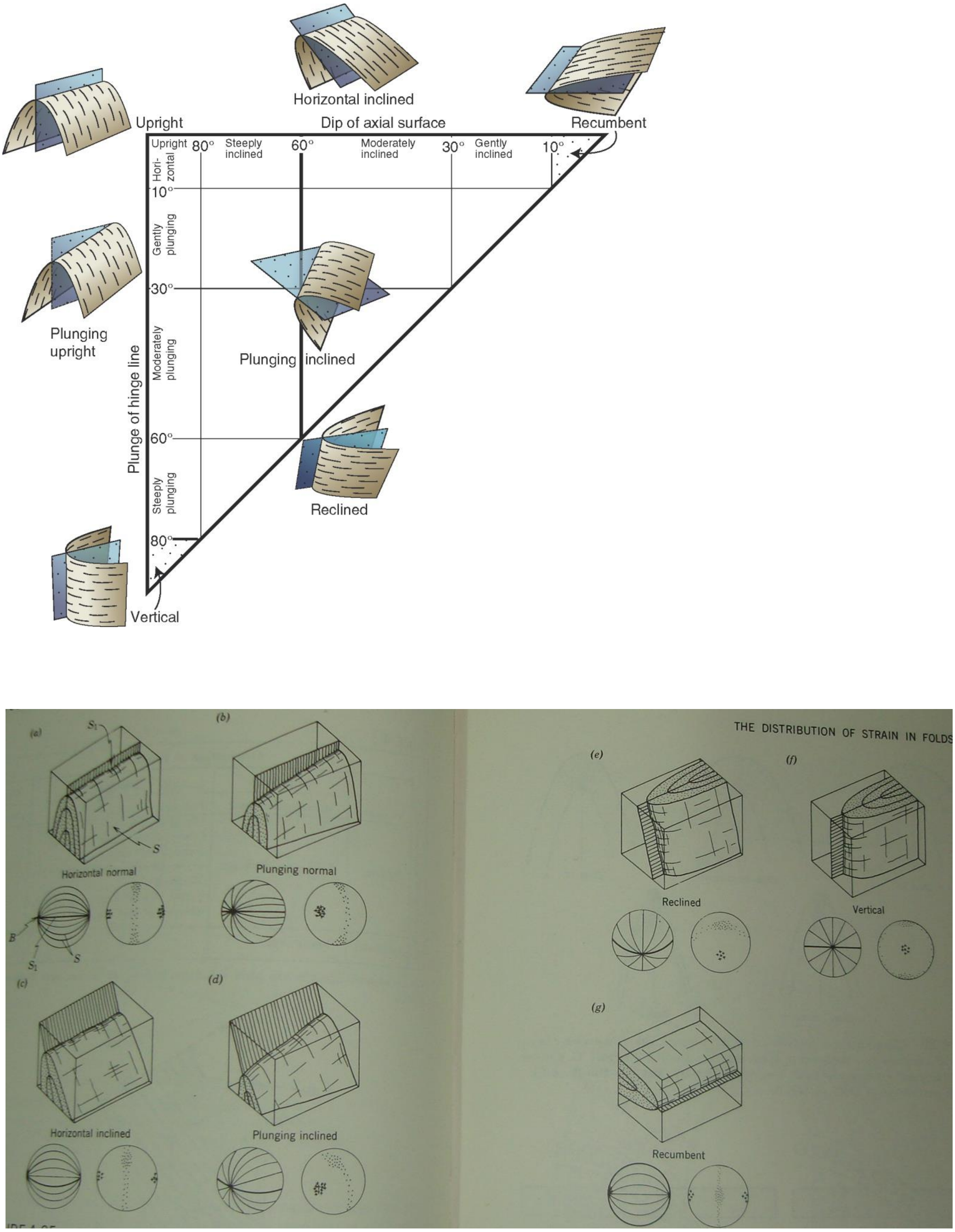


Fig.1.24.Shows the Classification of folds based on the orientation of axial surface and the orientation of hinge line.

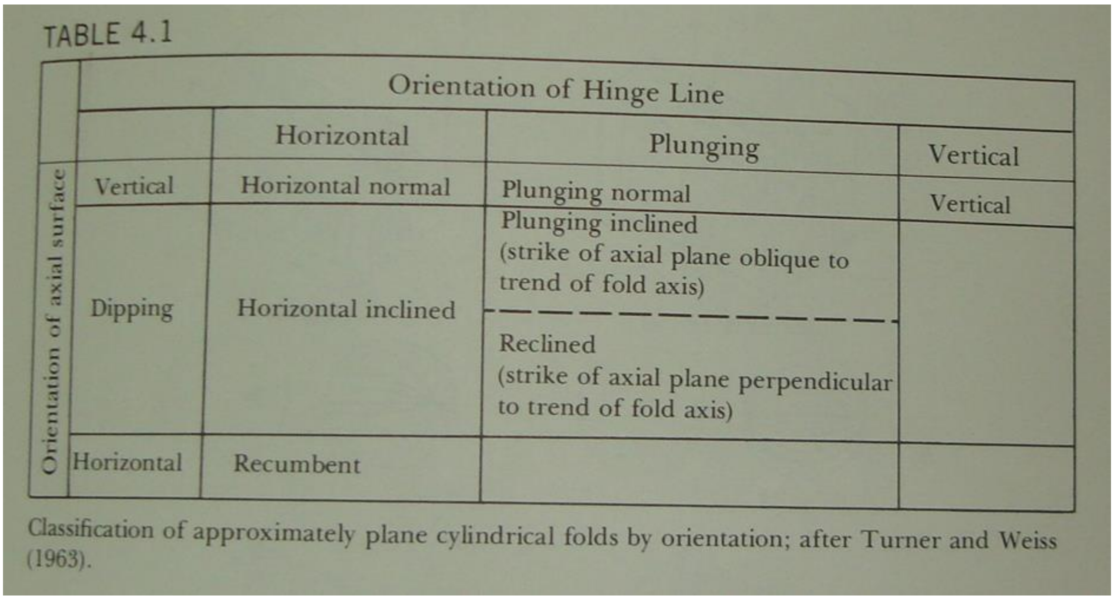


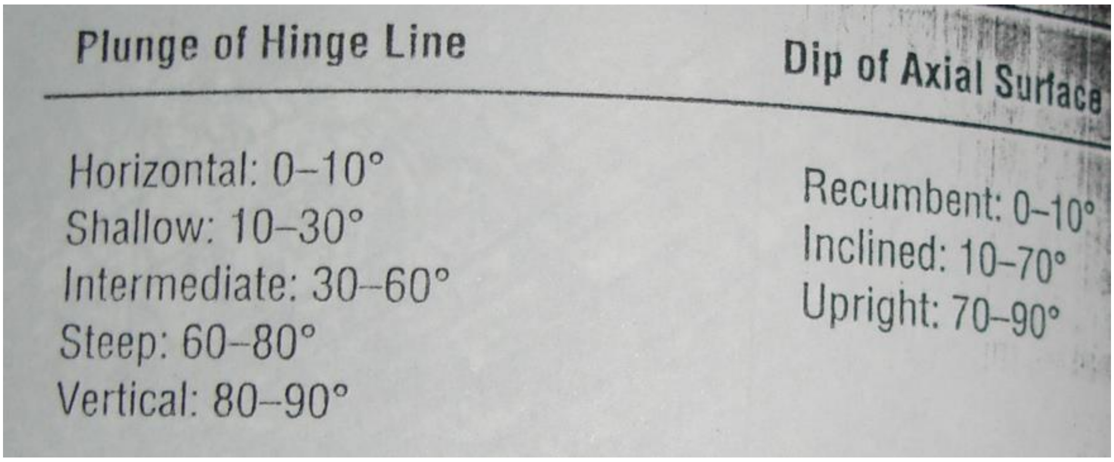
TABLE 4.1

Orientation of axial surface		Orientation of Hinge Line		
		Horizontal	Plunging	Vertical
	Vertical	Horizontal normal	Plunging normal	Vertical
	Dipping	Horizontal inclined	Plunging inclined (strike of axial plane oblique to trend of fold axis) Reclined (strike of axial plane perpendicular to trend of fold axis)	
	Horizontal	Recumbent		

Classification of approximately plane cylindrical folds by orientation; after Turner and Weiss (1963).

Table 2.

Depending on the angle of dipping of the axial surface, it's classified to three types of folds as **recumbent**, **reclined** and **vertical** or **upright** and depending on the angle of plunging of the hinge line, it's classified to five types, as **horizontal**, **shallow**, **intermediate**, **steep** and **vertical**, table 3.



Plunge of Hinge Line	Dip of Axial Surface
Horizontal: 0–10°	Recumbent: 0–10°
Shallow: 10–30°	Inclined: 10–70°
Intermediate: 30–60°	Upright: 70–90°
Steep: 60–80°	
Vertical: 80–90°	

Table.3

Mechanics and causes of Folding

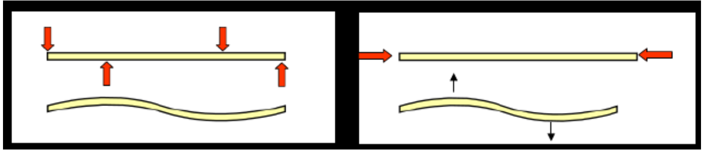
In general three types of folding are recognized, but the transition and combination are common. The types are (1) flexure folding, (2) shear folding, (3) flow folding

Kinematic analysis is the reconstruction of movements that take place during the deformation of rocks at all scales.

Folds may develop by a range of different mechanisms. These different mechanisms give rise to different deformation paths for different parts of the fold. Understanding the mechanism may enable you to predict the position of localization deformation (e.g. **hinge**, **limb** etc) and therefore may be useful in exploration.

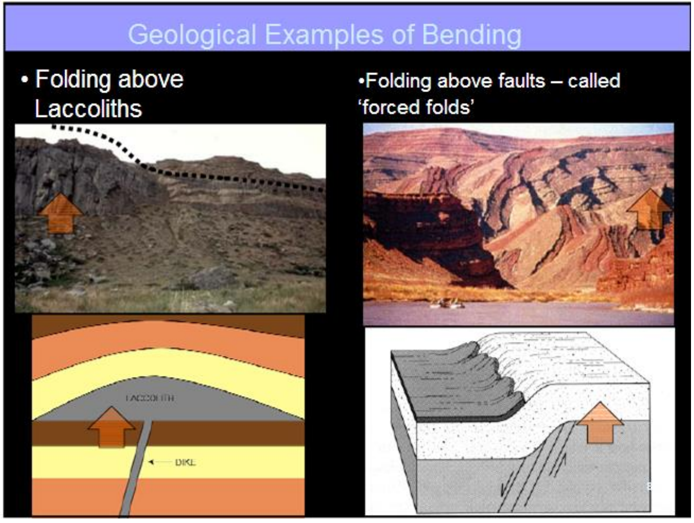
1- **Flexural Folding** : can result from bending or buckling of single layer or multiple layers . Bending & buckling describe the different ways forces are applied to a layer.

A: Bending: Layer is folded by deflection in same direction as applied stress, equal & opposite torques bend layer into a fold shape Fig.1.25-a. In pure bending, there is no net tension or compression averaged over the layer, either parallel or perpendicular to it. The vertical force acting upward in the middle of the layer could represent the uniform fluid pressure of an intrusive magma body applied over a segment of strata. The resulting uplift of the strata into a localized fold is called a **laccolith**. This vertical force could also represent the effect of a normal fault in basement rocks that bends the overlying strata into a **monoclinal fold** Fig1.26.



a b

Fig.1.25, (a) **Bending: require opposite torques,** **(b) Buckling: Compressive stress acting parallel to the layer**



Geological Examples of Bending

- Folding above Laccoliths

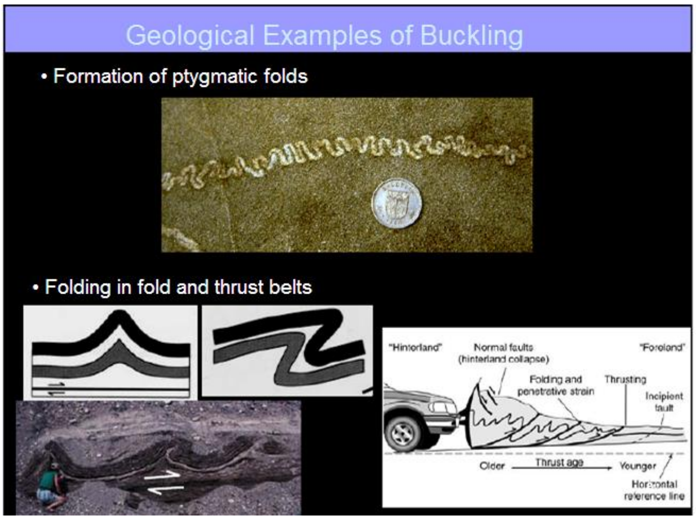
•Folding above faults – called 'forced folds'

Fig.1.26

B: Buckling: result from the application of compressive stress **parallel** to a competent layer Fig1.25- b. Then the layer is folded by deflection normal to the shortening direction.

Net compression causes layer to buckle into a fold shape. Buckling applies to a single folded layer of finite thickness or to multiple layers with high cohesive strength between layers.

Buckling is what happens when you push on the ends of fairly rigid layer (put a piece of paper on a flat surface and push the edges towards one another) . The compressive force here is parallel to the bedding Fig.1.27. Buckling of a layer will produce parallel fold (class 1b) geometries Fig.1.28, since the thickness of the layer is unaffected. The important thing to realize is that the model predicts a characteristic pattern of strain.



Geological Examples of Buckling

- Formation of ptygmatic folds
- Folding in fold and thrust belts

Fig.1.27

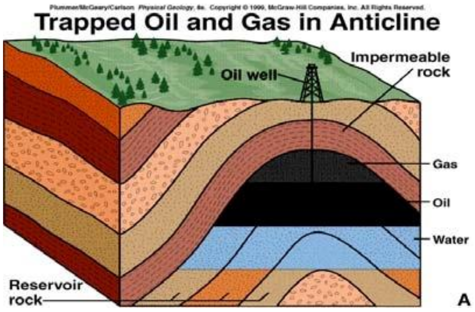


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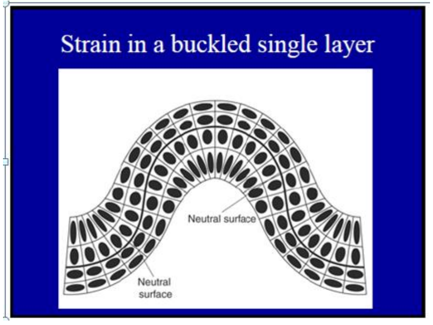
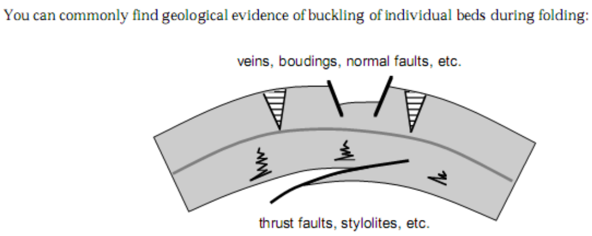
Fig.1.28. Shows trapped oil and gas in anticline, the fold is class1B.

Characteristic features of buckle folds: (1) the upper part of the layer folded anticlinally will be in extension, the lower half in compression. You can define a **neutral surface** that separates areas of compression and extension. On this surface, material points experience no strain. (2) Deformation occurs only by bending about the fold axis. Ideally, there is no extension parallel to the fold axis. That is, this is an example of **plane strain**. (3) Compressive and extensional strain increase with distance from the neutral surface. (4) There is no strain at inflection points of the limbs Fig.1.29. In the field we can show some indications of the buckling like the veins, normal faults, and others extension index on the convex side of the fold, and stylolites and reverse faults in its concave side Fig.1.30.



Strain in a buckled single layer



You can commonly find geological evidence of buckling of individual beds during folding:

Fig.1.29

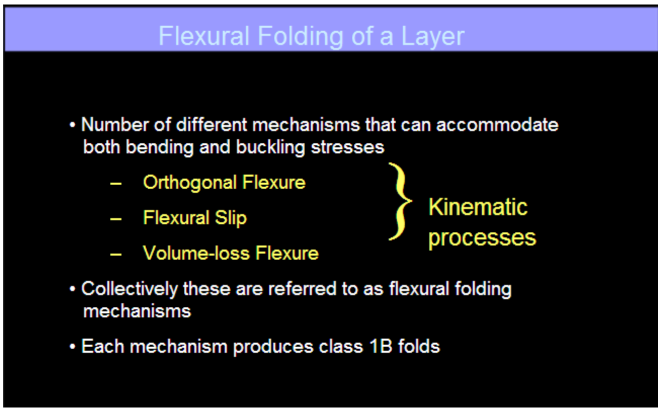
Fig.1.30

In the folding of sedimentary rocks, some formations are **competent**, whereas others are **incompetent**. Competency is a relative property. A **competence formation** is a strong and can transmit the compressive force much farther than a weak, **incompetent formation.** Many factors determine the competency of the rock, (1) crushing strength, the greater crushing strength will be the more competent in folding, as quartzite and marble are more competence than sandstone and limestone, and the shale is the weakest, (2) massiveness of the formation, in the same kind of rock (e.g. limestone), the thicker beds is more competent, (3) healing of fractures, the sandstone may be stronger than an adjacent limestone. But once the sandstone has broken, the fracture may heal with difficulty, whereas the rupture in the limestone may heal relatively rapidly, so the limestone is more competent than the sandstone.

C Passive folding, If layer rheology does not control the folding process, it is passive folding.

Kinematic Model:

A layer may respond to either bending or buckling load by orthogonal flexure or flexural slip or volume-loss flexure which are kinematic processes.



Flexural Folding of a Layer

- Number of different mechanisms that can accommodate both bending and buckling stresses
  - Orthogonal Flexure
  - Flexural Slip
  - Volume-loss Flexure

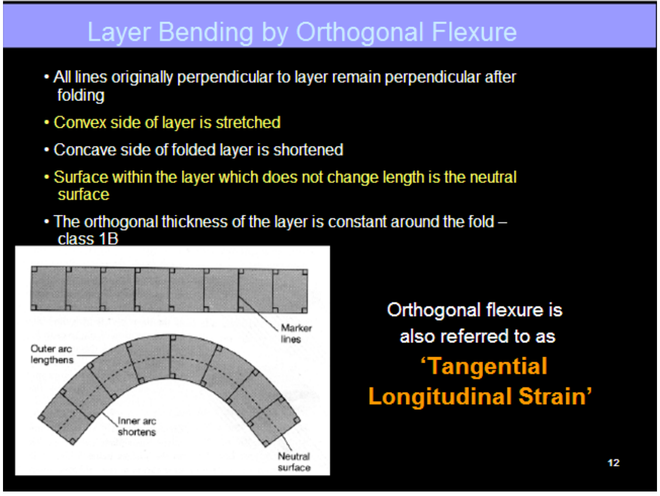
Kinematic processes

- Collectively these are referred to as flexural folding mechanisms
- Each mechanism produces class 1B folds

Orthogonal Flexure

Strain in buckled single layer

This type of mechanism has been covered in the paragraph above as in the buckling.



Layer Bending by Orthogonal Flexure

- All lines originally perpendicular to layer remain perpendicular after folding
- Convex side of layer is stretched
- Concave side of folded layer is shortened
- Surface within the layer which does not change length is the neutral surface
- The orthogonal thickness of the layer is constant around the fold – class 1B

Orthogonal flexure is also referred to as

'Tangential Longitudinal Strain'

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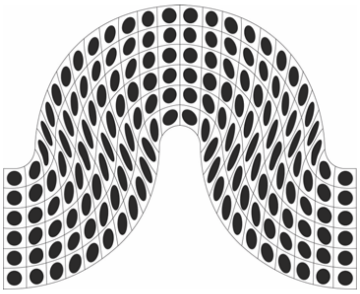
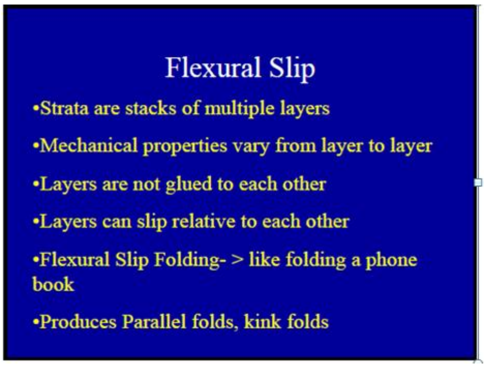
Flexural slip (shear)

This is “phonebook folding” or “deck of cards” folding. The idea is that folds are produced by shear on surfaces parallel to the layer being folded. This model produces **parallel folds**.

Important features are: 1) Deformation occurs by bending about the fold axis and shear on the slip surfaces in directions normal to the fold axis. It is also plane strain. Fold axis is parallel to the intermediate principal strain axis. 2) Layer maintains its thickness, but there is **no** neutral surface. 3) Folded layer is a circular section through strain ellipses. 4) Profile of fold shows a **divergent fan** of strain ellipses. 5) A passive linear marker on the fold surface will maintain its orientation Figs.1.31-1.32-and1.33.

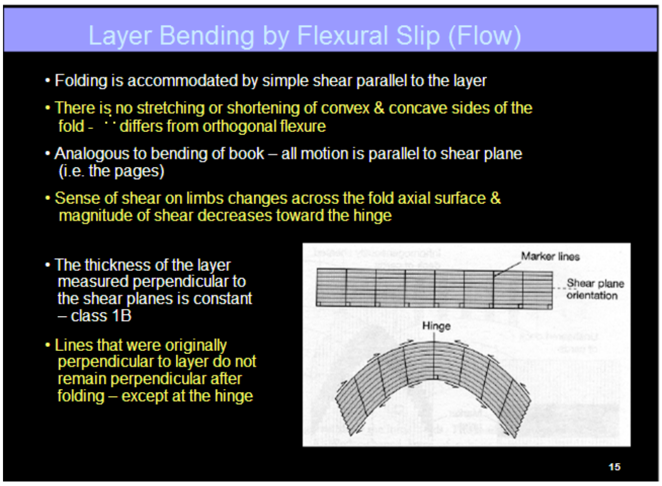




Flexural Slip

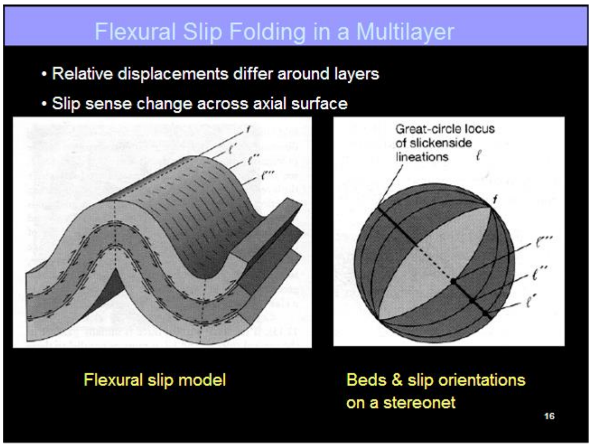
- Strata are stacks of multiple layers
- Mechanical properties vary from layer to layer
- Layers are not glued to each other
- Layers can slip relative to each other
- Flexural Slip Folding- > like folding a phone book
- Produces Parallel folds, kink folds

Fig.1.31. Flexural slip-folding multilayer



Layer Bending by Flexural Slip (Flow)

- Folding is accommodated by simple shear parallel to the layer
- There is no stretching or shortening of convex & concave sides of the fold - ∵ differs from orthogonal flexure
- Analogous to bending of book – all motion is parallel to shear plane (i.e. the pages)
- Sense of shear on limbs changes across the fold axial surface & magnitude of shear decreases toward the hinge
- The thickness of the layer measured perpendicular to the shear planes is constant – class 1B
- Lines that were originally perpendicular to layer do not remain perpendicular after folding – except at the hinge



Flexural Slip Folding in a Multilayer

- Relative displacements differ around layers
- Slip sense change across axial surface

Flexural slip model

Beds & slip orientations on a stereonet

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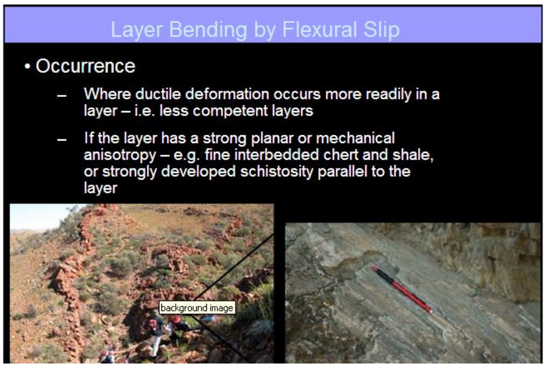
Fig.1.32.

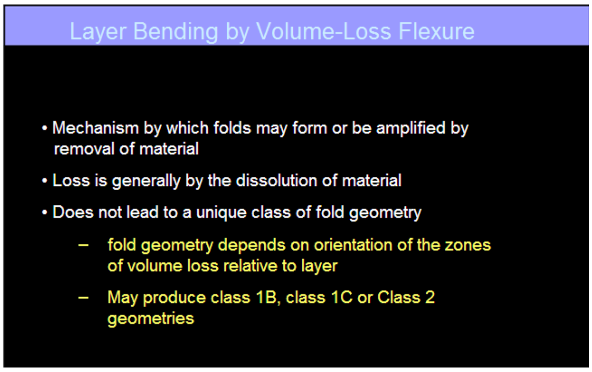


Layer Bending by Flexural Slip

- Occurrence
  - Where ductile deformation occurs more readily in a layer – i.e. less competent layers
  - If the layer has a strong planar or mechanical anisotropy – e.g. fine interbedded chert and shale, or strongly developed schistosity parallel to the layer

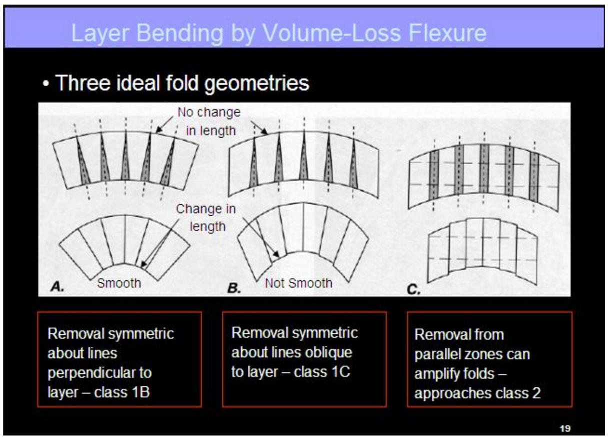
Fig.1.33.

Volume-loss flexure is a process by which folds are created or amplified by the dissolution of particular zones of the folded layer Fig.1.34; this process may also be called *solution folding*. Because the final form of the fold depends on where material is removed, this kinematic model does not produce single resultant fold geometry (class). It may also produce the illusion that shearing has occurred Fig.1.35, though none necessarily has when homogeneous flattening may be combined with the processes above to create new fold geometries. For example, a class 1B fold, when subject to flattening perpendicular to the hinge, is transformed in to a class 1A fold. However, a class 2 fold remains such when subject to flattening.



Layer Bending by Volume-Loss Flexure

- Mechanism by which folds may form or be amplified by removal of material
- Loss is generally by the dissolution of material
- Does not lead to a unique class of fold geometry
  - fold geometry depends on orientation of the zones of volume loss relative to layer
  - May produce class 1B, class 1C or Class 2 geometries


Layer Bending by Volume-Loss Flexure

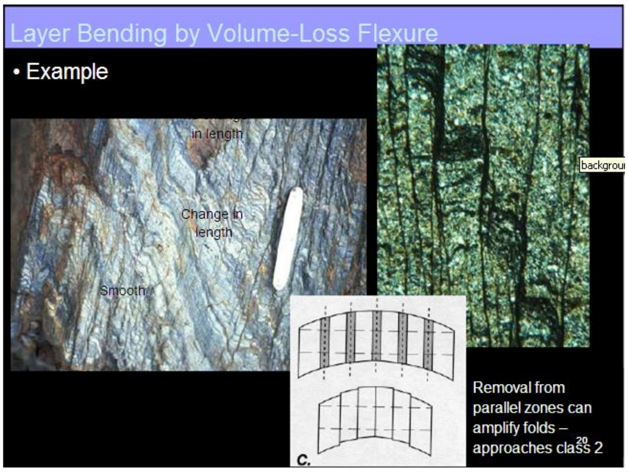
- Three ideal fold geometries

Removal symmetric about lines perpendicular to layer – class 1B

Removal symmetric about lines oblique to layer – class 1C

Removal from parallel zones can amplify folds – approaches class 2

Fig.1.34.


Layer Bending by Volume-Loss Flexure

- Example

Removal from parallel zones can amplify folds – approaches class 2

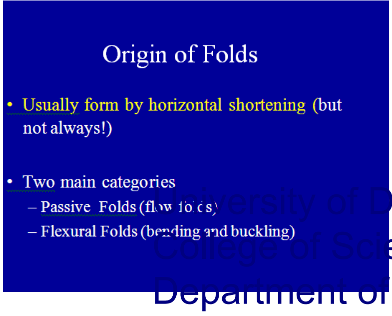
Fig.1.35.

Passive Shear folding

Typically occurs in layers of low competence and takes place by inhomogeneous simple shear Fig.1.36; the fold axis and hinge are not related to the direction of shear, and the model permits no shortening normal to the shear planes. This results in class 2 (*similar*) folds.

Passive shear folding also called passive flow folding or simply flow folding, occur in an incompetent layer, which acts simply as a marker that record the deformation Fig.1.37.

The layer being folded is assumed to exert no mechanical effect on the folding process. It is just a passive marker. Folds form by differential flow or slip along closely spaced surfaces oblique to the layer being folded. This will produce class2 fold or **similar folds** Fig.1.38.


Origin of Folds

- Usually form by horizontal shortening (but not always!)
- Two main categories
  - Passive Folds (flow folds)
  - Flexural Folds (bending and buckling)

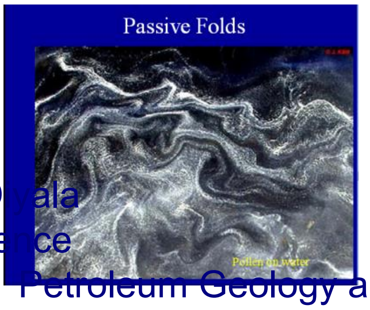
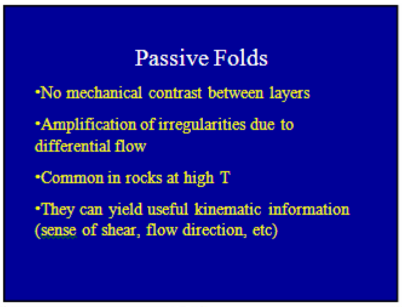
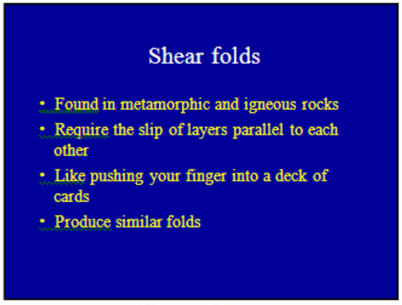

Passive Folds

Fig.1.36



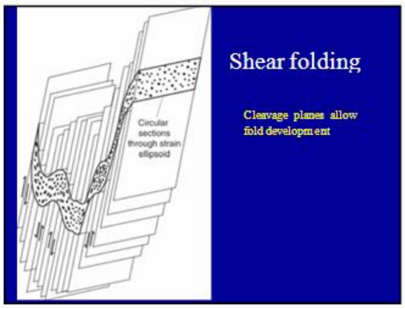
Passive Folds

- No mechanical contrast between layers
- Amplification of irregularities due to differential flow
- Common in rocks at high T
- They can yield useful kinematic information (sense of shear, flow direction, etc)



Shear folds

- Found in metamorphic and igneous rocks
- Require the slip of layers parallel to each other
- Like pushing your finger into a deck of cards
- Produce similar folds



Shear folding

Cleavage planes allow fold development

Shear folding

Crystallina fold- Alps

J. Amato©

Fig1.37. Fig.1.38

Characteristic features: 1) Deformation by simple shear on shear planes. The shear planes are circular sections through the strain ellipsoid. 2) The direction of shearing can be quite variable. The only requirement is that shear plane is not parallel to the layer. However, the maximum fold amplitude obtains when shear direction is normal to the fold axis. 3) In the plane parallel to the axial plane of the fold, no changes in layer thickness will be observed. However, viewing the fold normal to the shear plane and parallel to the presumed shear direction, you will see great variation between the thickness of hinges and limbs. This variation requires no flow of material from limbs to hinges. 4) Shear sense on the shear planes changes from limb to limb on the same fold. Strain ellipses, and principal planes of strain make a divergent fan. 5) There is no neutral layer, and strains are constant at all points within a layer. 6) These folds can be harmonic over large lengths, as opposed to concentric folds, where geometry requires them to be detached. 7) A passive linear marker is distorted and rotates towards the slip direction. Since the slip direction is never parallel to the hinge of the fold produced, initially linear markers will never be oriented parallel to the fold hinge. 8) There is no relationship between layer thickness and wavelength of folds.

Brittle Deformation

Brittle deformation is simply the permanent change that occurs in a solid material (rocks) due to the growth of fractures and/ or sliding on fractures once they have formed. Brittle deformation occurs only when stresses exceed a critical value, and thus only after a rock has already undergone some elastic and/or plastic behavior.

Fracture: is a planar or curviplanar discontinuity forms as a result of brittle rock failure under relatively low pressure and temperature condition in the earth crust. These structural discontinuities are amongst the most common of all geological features: every outcrop and most cores exhibit some sort of fracturing. Fractures and other discontinuities affect nearly every **petroleum reservoir**, either by enhancing the production, or by causing problems for production.

Rock fractures range in **size** from microcracks (fraction of mm) to faults which extend for hundreds of kilometers.

Fracture is a discontinuity across which **cohesion** (C_o) is lost.

A fracture dose not extend infinitely in all direction, some fracture intersect the surface of a body of rock, whereas others terminate within the body. The line representing the intersection of the fracture with the surface of a rock body is the **fracture trace** and the separating the region of the rock that has fractured from nonfractured regions is the **fracture front.** The point at which the fracture trace terminate on the surface of the rock is the **fracture tip, Fig(3-1).**

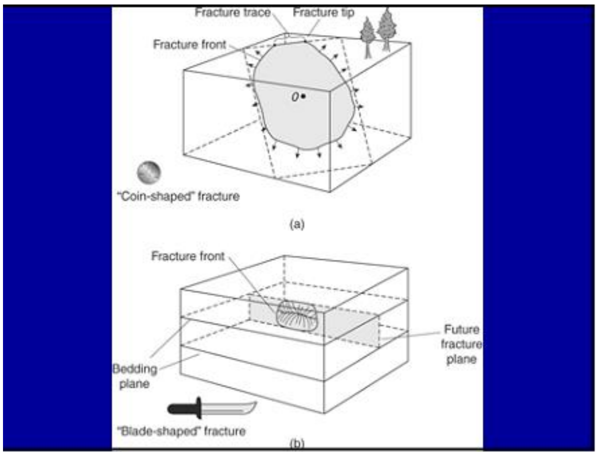


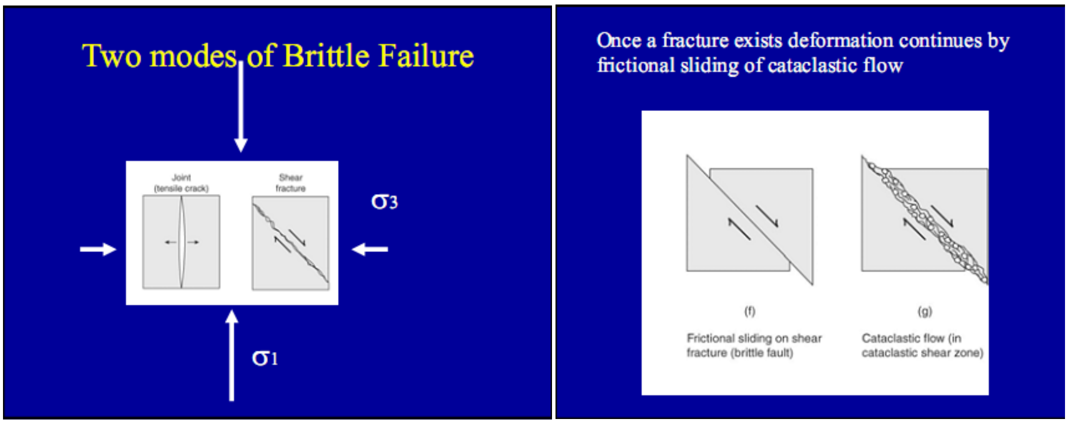
Fig. (3-1)

The term fracture **includes** three basic types of discontinuities:

Extension fracture (type I): Relative movement normal to fracture surface. Tensile cracks form normal to the σ_3 (parallel to the $\sigma_1\sigma_2$ plane)

Crack opens infinitesimally perpendicular to the crack plane (Fig.3-2)

Crack grows in its own plane; no bending/changing orientation.



Two modes of Brittle Failure

Once a fracture exists deformation continues by frictional sliding of cataclastic flow

Fig. (3-2)

Shear fracture (type II & III): Relative movement parallel to fracture surface.

Mode II – Sliding Mode: One block moves parallel to the fracture surface and normal to the fracture front or tip.

Mode III – Tearing Mode: One block moves parallel to the crack and parallel to the fracture front or tip (Fig 3-3).

Oblique extension (hybrid) fracture: A fracture that has components of displacement both perpendicular and parallel to fracture surface.

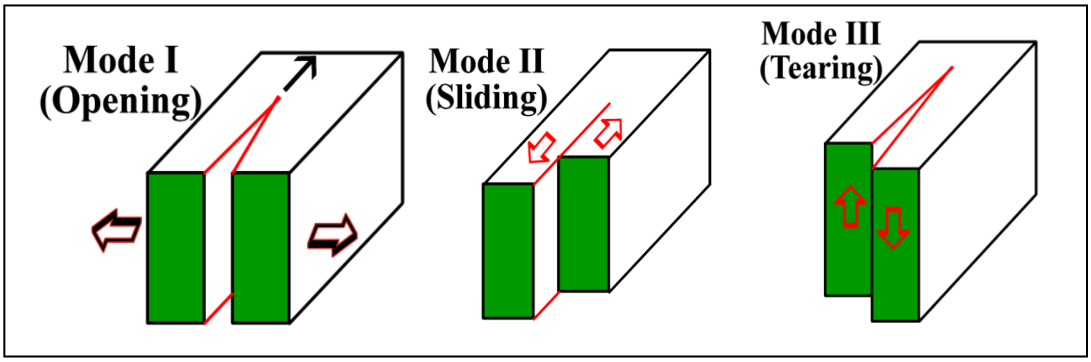


Fig. (3-3)

Wing cracks: a tensile stress concentration occurs at the ends of a crack that is being loaded by a shear stress.

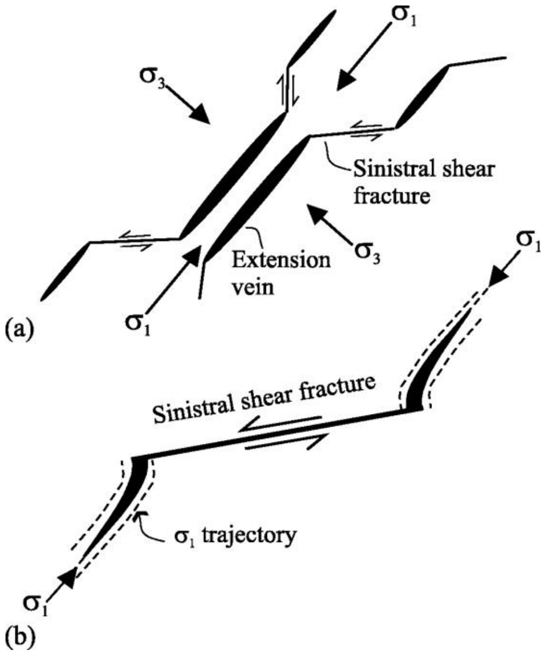
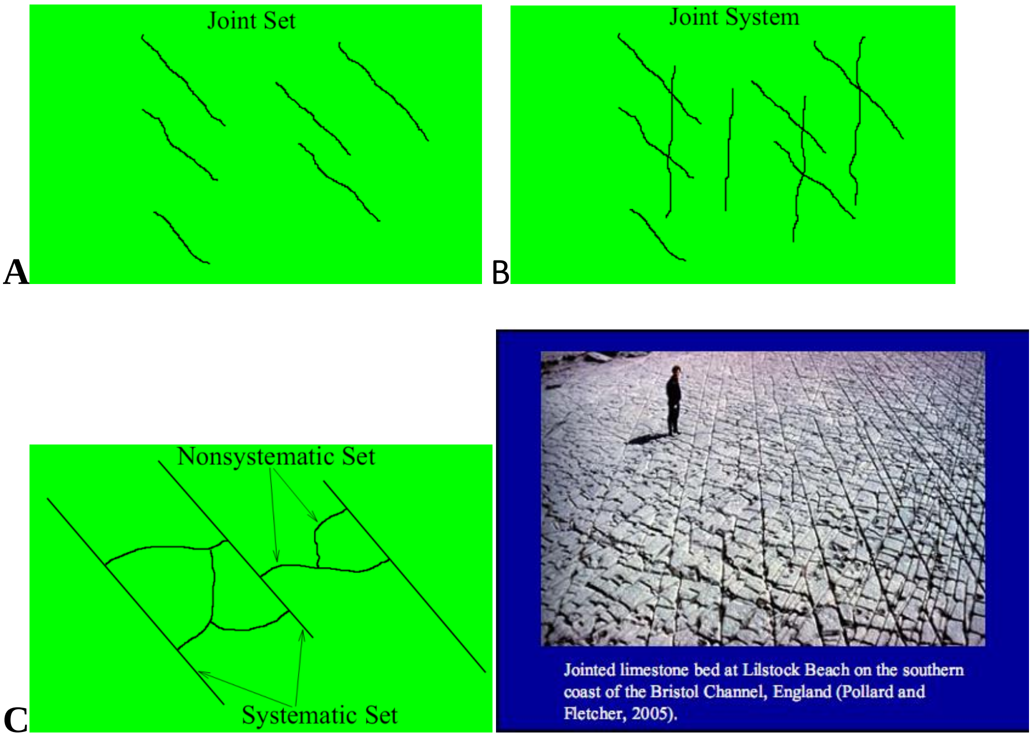


Fig. 3-3b shows wing crack

Fracture set and system:

- **Fracture or joint sets:** a group of fractures occurs in systematic alignment with similar strike and dip and arrangement (Fig.3-4A)**.**
- Small extension fractures are referred to as joint.
- **Systematic joints:** have roughly planar fracture surfaces, regular parallel orientation and regular spacing (Fig.3-4C).
- **Non-systematic joints**: are curved and irregular in geometry, although they are distinguished by nearly always terminating against older joints that belong to a systematic set(Fig.3-4C).
- **Fracture system (fracture assemblages):** two or more sets of fracture affecting the same volume of rock each set being characterized by a different strike and dip (Fig.3-4B).

JOINTS: are defined as dry fractures of geologic origin along which no appreciable displacement has occurred. In sedimentary rocks these joints are usually perpendicular or parallel to the bedding plane. In volcanic rocks, the contraction of the rock during cooling forms joints that isolate prisms perpendicular to the gradient of temperature. In granitic rocks, joints appear to be related to the relaxation of the vertical stress during erosion and exhumation.



Jointed limestone bed at Lilstock Beach on the southern coast of the Bristol Channel, England (Pollard and Fletcher, 2005).

Fig. 3-4

Why are the Joints important?

- Provide a mass wasting surface failure plane; joint analysis typically done for slope stability, dam stability, tunnel stability.
- Produce strength anisotropy: later reactivation?
- Provide fracture porosity/permeability - hydrologic modeling, mineralization.
- Important geomorphic control, contributing to drainage (e.g. trellis), lineaments.
- At deeper levels joints exert a control on the migration of geological fluids: water, petroleum and gas.
- They provide an easy to interpret paleostress system (with caution), but difficult to date (igneous rocks provide an exception to difficulty in dating because of late juices that can fill the joints to produce veins).

Some common joint associations

In different geologic situations you get different characteristic joint patterns.

- joints in volcanic rocks.
- joints in plutonic rocks related to thermal contraction.
- joints associated with folds and tectonism: usually at least two sets in symmetry, but up to 5.
- intracratonic regional joint sets.
- joint sets associated with point phenomena and radial and concentric patterns.
- conjugate sets: two sets at roughly 60 degrees to each other that form together (why 60?).

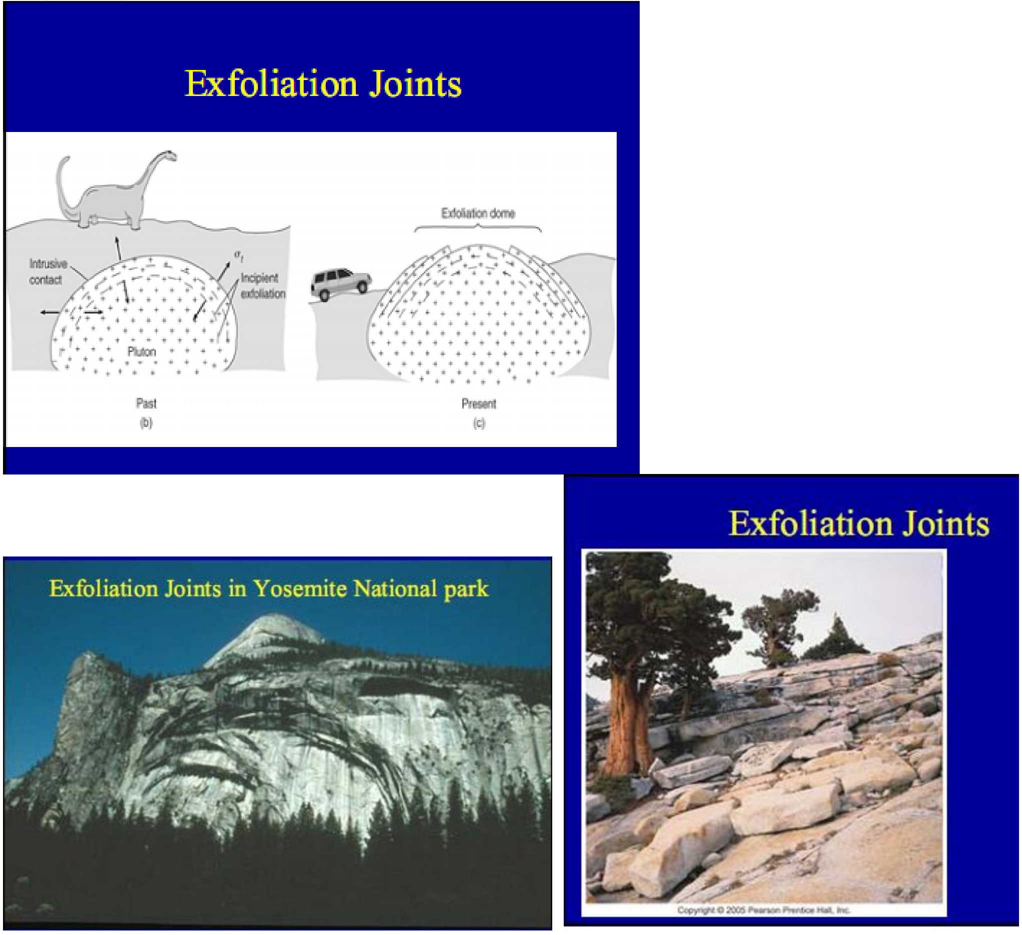
Joints are sometimes farther classified as **extension joints** and **shear joints**, as a subdivision based on the angular relations of crossing joints. Because no movement normal or parallel to the joint walls can be observed. Although individual joint fractures may be quite short (1-10m), in certain regions it is found that master joints run for very long distances. Many of the striking lineaments seen on air photographs are master joints rather than the major faults.

Thus, the main types of joint are:
Tectonic joints; breaks formed from the tensile stresses accompanying uplift or lateral stretching, or from the effects of regional tectonic compression. They commonly occur as planar, rough-surfaced sets of intersecting joints, with one or two of the sets usually dominating in persistence (Fig.3-5).

Fig. 3-5 Note how the rock is broken into rectangular blocks, a common phenomenon when two orthogonal (at right angles) joint sets exist. Also note how the one set of joints truncates against the continuous joint running from side to side in the image. If this relationship is consistent, the continuous joint is the longitudinal joint, which is interpreted to have formed **before** the cross joints that truncate against it. Finally note that some fractures that can not be assigned to one of the two orthogonal sets exist. It is not uncommon at all that a rock body can have three or more joint sets developed within it.

Sheeting joints or Exfoliation joints; a set of joints developed more or less parallel to the surface of the ground, especially in plutonic igneous intrusions such as granite; probably as a result of the unloading of the rock mass when the cover is eroded away. It appears that sheeting joints form where horizontal stress is greater than the vertical load.

Breaks developed as a product of exfoliation; the breaking or splitting off from bare rock surfaces by the action of chemical or physical forces, such as differential expansion and contracting during heating and cooling over the daily temperature range (Fig. 3-6).


Exfoliation Joints

Exfoliation Joints in Yosemite National park

Exfoliation Joints

Fig. (3-6). Shows sheeting and exfoliation joints.

Cooling joints, are extensional fractures characteristic of shallow tabular igneous intrusion, dykes or sills or thick extrusive flows. The fractures separate the rock into roughly hexagonal or pentagonal columns, which are often oriented perpendicular to the contact of the igneous body with the surrounding rock (Fig.3-7).

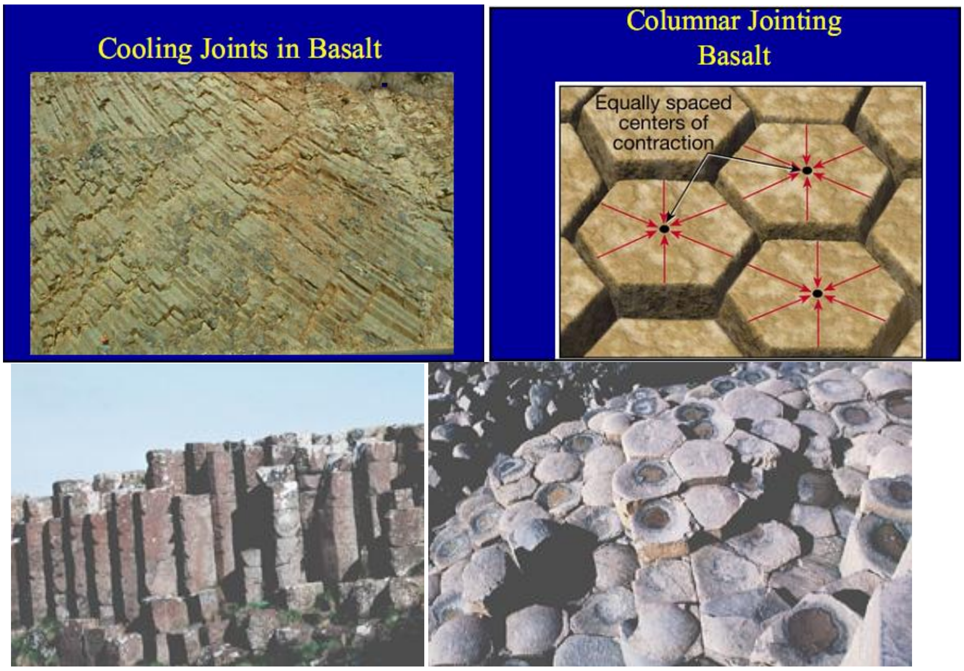


Fig.(3-7)

Pinnate fractures or **feather fractures** are extension fractures that form en echelon arrays along brittle shear fractures, The **acute angle** between the feather joints and the fault points in the direction of relative movement of the block on which it lies. They may form before or during the faulting process and have a relationship to the regional stress field that is the same as that for **en echelon** extension fractures, which means the extension fractures are parallel to one another but offset from one another along the trend of the shear fracture, which is oblique to the extension fracture plane. The sense of rotation through the acute angle from the fault plane to the extension fracture plane is the same as the shear sense on the fault plane (Fig.3-8).

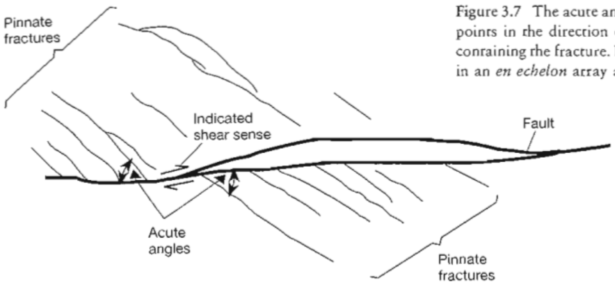


Figure 3.7 The acute angle between the fracture and the fault points in the direction of relative motion of the fault block containing the fracture. Pinnate fractures, or feather fractures, in an *en echelon* array along a brittle fault.

Fig. (3-8)

Gash Fractures: are extension fractures usually mineral-filled S- or Z-shaped, that form **en echelon** sets as extensional fractures formed in **ductile** shear zones. The orientation of gash fractures relative to the shear zone can be used to determine the sense of shear on the associated shear zone (Fig.3-9).

Fig.(3-9)

Features of Fracture Surfaces

Features on the surface of a fracture provide information about the fracture's origin. A **plumose structure**, or **hackle plume**, has a characteristic feather pattern. **Arrest line** are curved features perpendicular to the lines of hackle of the fracture face. All these features indicate *extensional* fracturing, as opposed to **slickenside lineation**, which indicate *shear* fracturing (Fig.3-10).

Plumose Structures are fracture networks that form at a range of scales, and spread outward from a *joint origin*. The joint origin represents a point at which the fracture begins. The *mirror zone* is the joint morphology closest to the origin that results in very smooth surfaces. *Mist zones* exist on the fringe of mirror zones and represent the zone where the joint surface slightly roughens. *Hackle zones* predominate after mist zones, where the joint surface begins to get fairly rough. **Hackle fringe** describes a set of extension fractures that are aligned *en echelon* and rotated away from the joint axis. This hackle zone severity designates *barbs*, which are the curves away from the *plume axis(Fig3-10*).

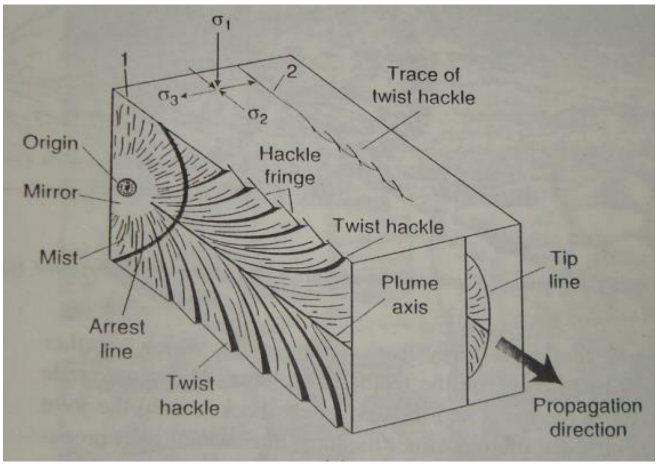
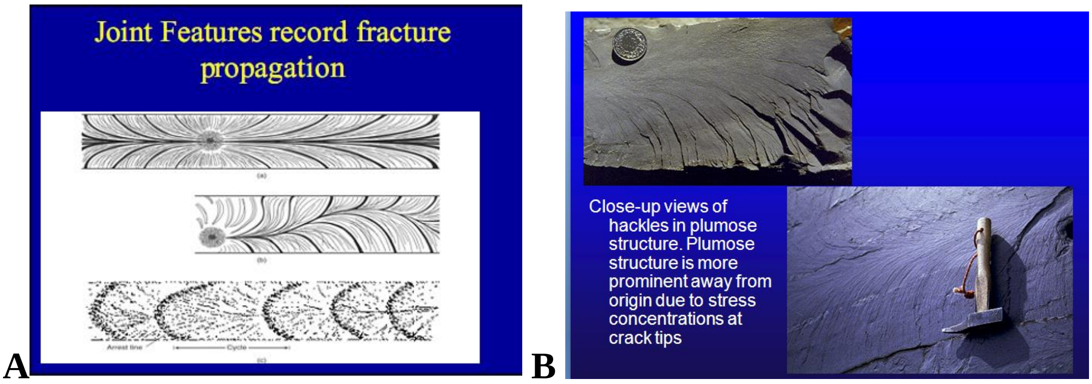


Fig.3-10 Shows the structure of plumose joint .



Joint Features record fracture propagation

Close-up views of hackles in plumose structure. Plumose structure is more prominent away from origin due to stress concentrations at crack tips

Fig.(3-11) Shows A, Joint features record fracture propagation, B, photos of plumose joints

Geometric classification of joints

Orthogonal Joints: occur when the joints within the system occur at mutually perpendicular angles to each other.

Two parameters influence **fracture patterns**: the **orientation** of the fractures and their **frequencies**. Orientation of fractures is based on the state of **stress** within the rock. In contrast, the frequency or spacing of fractures is based on the **properties** of the rocks in which the fractures have formed.

Three mutually perpendicular directions depend on the orthogonal axes **a**, **b** and **c**. At each locality the bedding defines the ab with the c-axis as normal, and with b-axis parallel to the fold axis and a-axis is parallel to dip direction. Joints parallel to the ac-plane are known as **cross joints**. And joints parallel to bc-plane as **longitudinal joints**, and those parallel to **ab** as **bedding joints**. At times, the orientation of the joint sets can be related directly to the folding and may be defined in terms of the ***a, b*** and ***c*** axes of the “tectonic cross” (Fig. 3.12). Those joints that cut the fold at right angles to the axis are called *ac* or cross joints. The *bc* or longitudinal joints are perpendicular to the latter joints, and diagonal or oblique joints run at an angle to both the *ac* and *bc* joints. Diagonal joints are classified as **shear joints**, whereas *ac* and *bc* joints are regarded as **tension** joints.

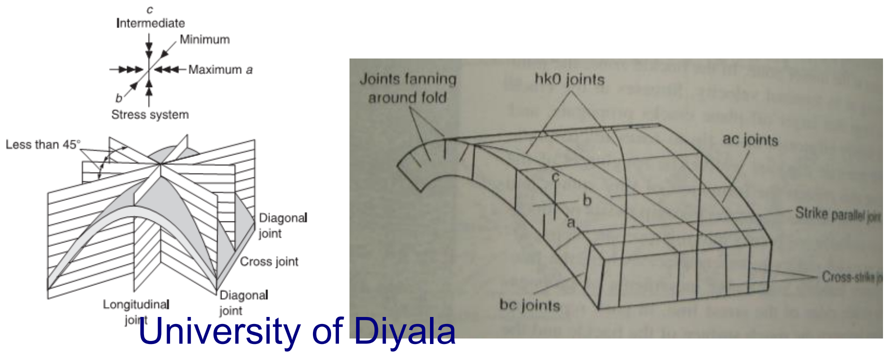


Fig.3-12 Geometric orientation of longitudinal, cross and diagonal joints relative to fold axis and to principal axes of stress.

Conjugate joints

This type of classification depend on the orientation of the joint with respect to the three axis (a, b, and c), when the joint crossing all the three axes is designated **hkl**, when it's parallel to a-axis as **Okl**, when it's parallel to b-axis as **hOl**, and when it's parallel to c-axis as **hkO** (Fig.3-13).

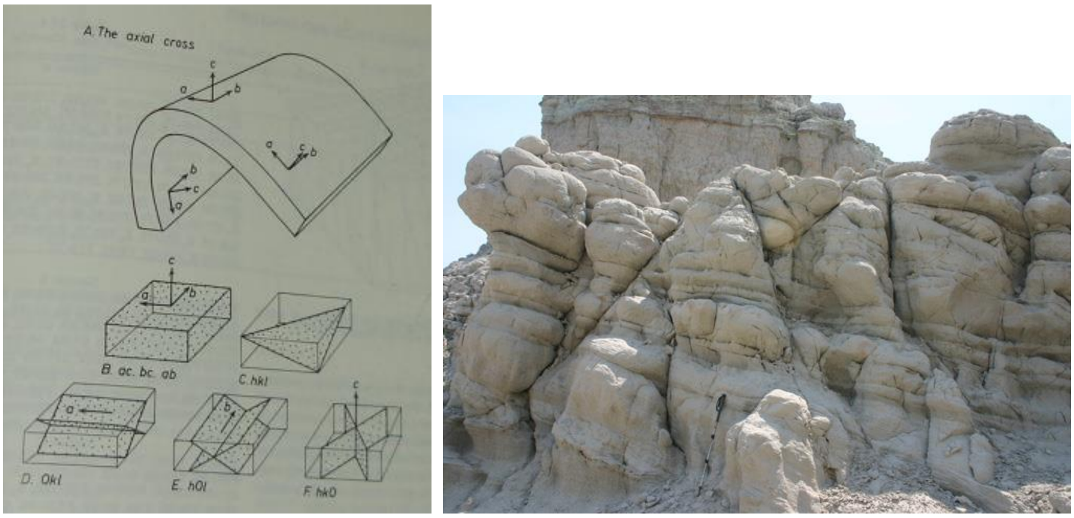


Fig 3-13.Example of conjugate joint set in tilted Tertiary sandstones of the Brule Formation in the Slim Buttes area of South Dakota. The interpretation is that the conjugate joint set happened before tilting of the strata.

Conjugate Joints occur when the joints intersect each other at angles significantly less than ninety degrees (Figs 3-14, 3-15, 3-16).

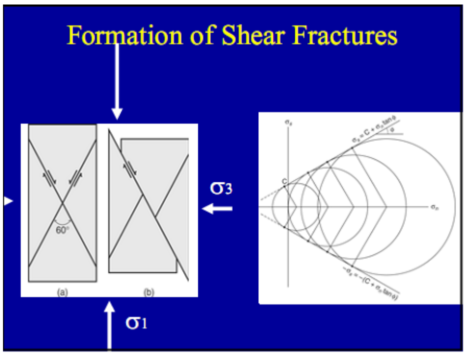

Formation of Shear Fractures

Fig. (3.14)

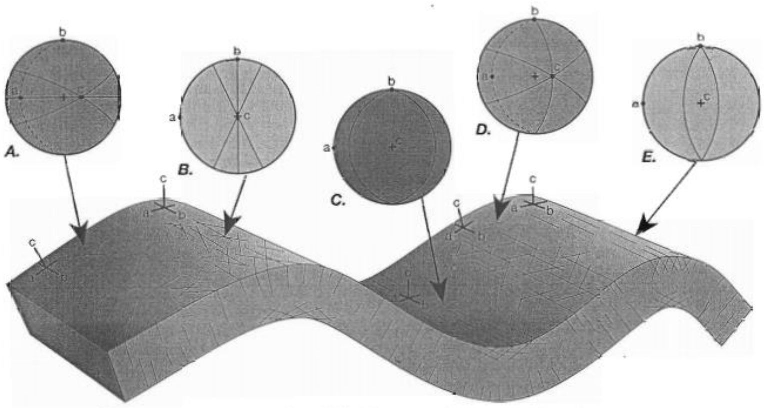
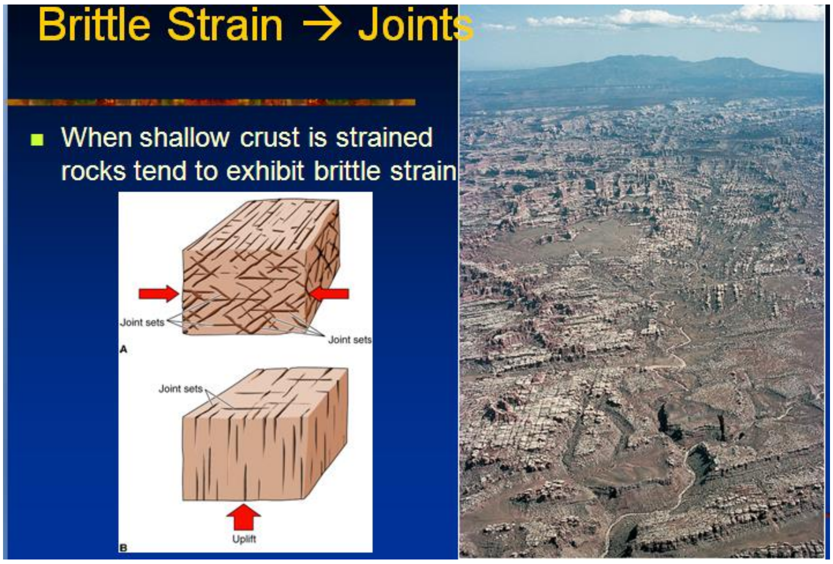


Figure 3.17 Fracrures associated wirh folds. The stereographic projections show the orieurations of the coordiuare system, the bedding where it is nor horizoutal (dotted great circles), and the fractures (solid great circles).

Fig. (3-15)



Brittle Strain → Joints

- When shallow crust is strained rocks tend to exhibit brittle strain

Fig.(3-16)

FAULT CLASSIFICATION AND TERMINALOGY

Faults: Are fractures that have appreciable movement parallel to their plane. They produced usually be seismic activity. Understanding faults is useful in design for long-term stability of **dams**, **bridges**, **buildings** and **power plants**. The study of fault helps understand mountain building. Faults may be hundred of meters or a few centimeters in length. Their outcrop may have as knife-sharp edges or fault shear zone. Fault shear zones may consist of a serious of interleaving anastomosing brittle faults and crushed rock or of ductile shear zones composed of mylonitic rocks.

What is a Fault?

- A fault is a break or fracture between two blocks of rocks in response to stress.
- Three types of stresses produce faults

1) Tension 2) Compression 3) Shear

- One block has moved relative to the other block. • The surface along which the blocks move is called a fault plane.

Parts of Faults: Fig. 4-1

A **fault line** is the surface trace of a fault, the line of intersection between the fault plane and the Earth's surface. Since faults do not usually consist of a single, clean fracture, geologists use the term **fault zone** when referring to the zone of complex deformation associated with the fault plane.

Fault scarp

The fault scarp is the feature on the surface of the earth that looks like a step caused by slip on the fault.

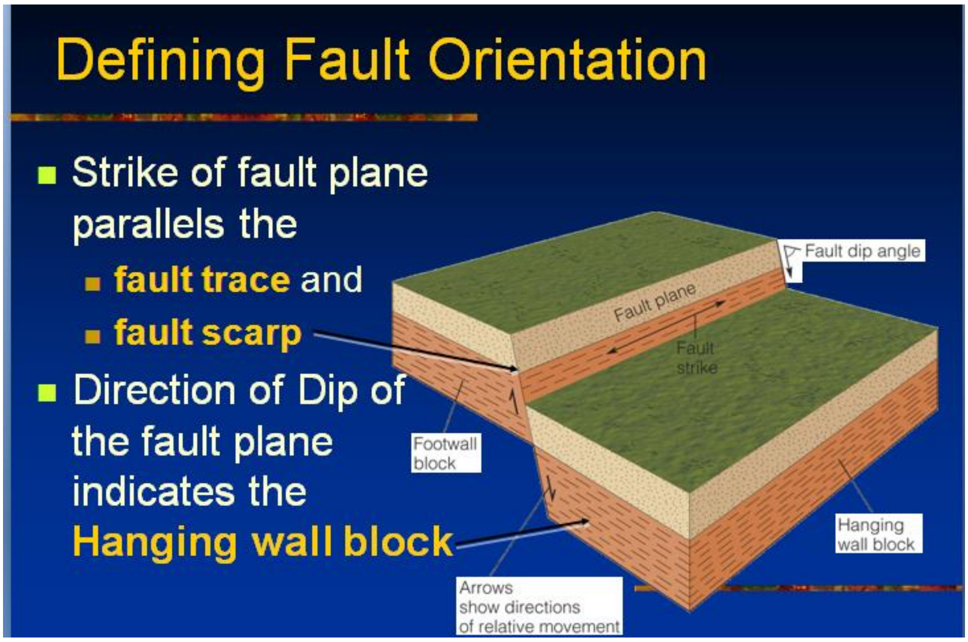
Fault trace

The fault trace is the intersection of a fault with the ground surface; also, the line commonly plotted on geologic maps to represent a fault.

Fault plane: Surface that the movement has taken place within the fault. On this surface the dip and strike of the fault is measured.

Hanging wall: The rock mass resting on the fault plane.

Footwall: The rock mass beneath the fault plane.



Defining Fault Orientation

- Strike of fault plane parallels the
  - **fault trace** and
  - **fault scarp**
- Direction of Dip of the fault plane indicates the **Hanging wall block**

Fig. (4-1)

Slip: Describes the movement parallel to the fault plane. **Dip slip:** Describes the up and down movement parallel to the dip direction of the fault. **Strike slip:** Applies where movement is parallel to strike of the fault plane. **Oblique slip:** Is a combination of strike slip and dip slip. **Net slip** (true displacement): Is the total amount of motion measured parallel to the direction of motion.

Separation: The amount of apparent offset of a faulted surface, measured in specified direction. There are **strike separation**, **dip separation**, and **net separation**.

Heave: The horizontal component of dip separation measured perpendicular to strike of the fault. **Throw:** The vertical component measured in vertical plane containing the dip (Fig. 4-2).

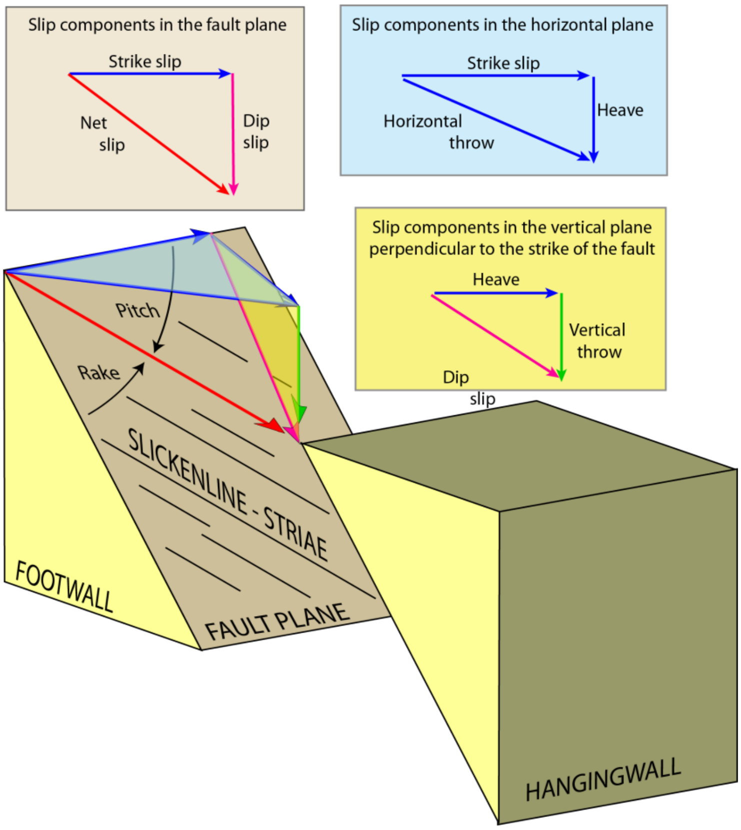


Fig. (4-2) Shows oblique slip fault accompanied by Slikenline-striae

Features on the fault surface are:
1) Grooves (parallel to the movement direction) 2) Growth of fibrous minerals (parallel to the movement direction) 3) Slickensides are the polished fault surfaces. 4) Small steps.
All are considered a kind of lineation. They indicate the movement relative trend NW, NE … etc.
Small steps may also be used to determine the movement direction and direction of movement of the opposing wall. Slicklines usually record only the last moment event on the fault.

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

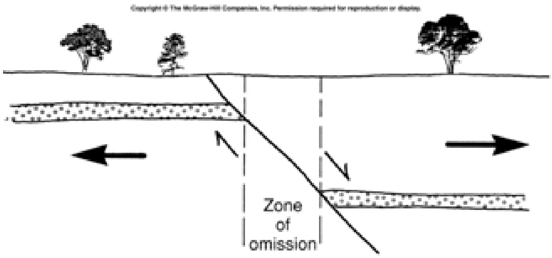
- a fault where the relative movement (or slip) on the fault plane is approximately vertical is known as a dip-slip fault (Fig 4-3).
- where the slip is approximately horizontal, the fault is known as a transcurrent, wrench or strike-slip fault (Fig. 4-3)

- an oblique-slip fault has non-zero components of both **strike and dip** slip.

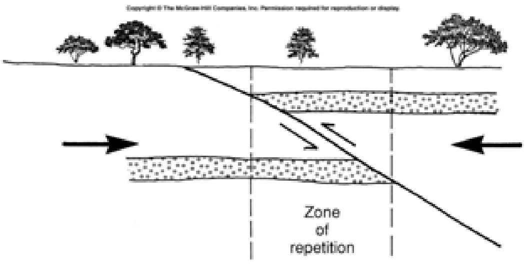
GEOLOGICAL CONCEPTS

TYPES OF FAULT

Normal and Reverse faults



Cross section of a normal fault.
Hanging wall moved downwards.
Zone of omission indicates
extensional forces active

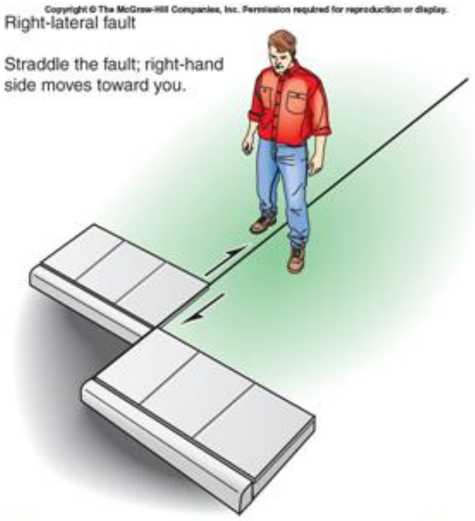


Cross section of a reverse fault.
Hanging wall has moved upwards.
Zone of repetition indicates
compressional forces active.

GEOLOGICAL CONCEPTS

TYPES OF FAULT

Strike-slip fault



The right hand side of the fault moves towards the man when he straddles the fault.

When he turns around the right hand side will still have moved closer to him.

Called a right lateral fault

Fig. 4-3 Shows types of faults

GEOLOLOGICAL CONCEPTS

TYPES OF FAULT

Right stepping right lateral fault

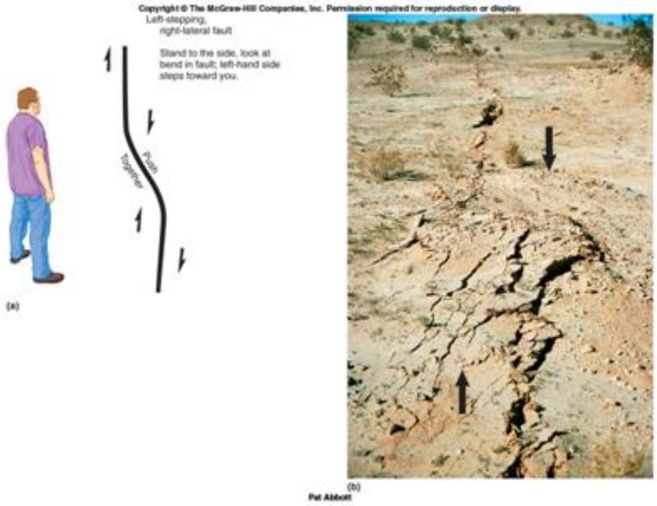
(a) The land is pulled apart at the fault bend whenever the fault moves. It creates a hole the could become a basin.

(b) Land offset along Superstition Hills right lateral fault during the rupture. Note the right step and the pull apart at the bend.

GEOLOGICAL CONCEPTS

TYPES OF FAULT

Left stepping right lateral fault



(a) A left step in a right lateral fault. The land is pushed together when the fault moves. Eventually could grow into a mountain.

(b) Example shows land offset along the Superstition Hills right lateral fault during the 16 November 1987 earthquake.

Fig. (4-4) Shows the concepts of left and right stepping in right lateral faults.

The attitude of the fault is important because it is used to classify the fault as either dip-or strike-slip. For example, if the displacement a cross the fault is parallel to strike, then the fault is a **strike-slip fault** (Figure 4-5).

On the other hand, if the displacement is parallel to the dip and at right angles to the strike, then the fault is a **dip-slip fault** (Figure 4- 5). However, sometimes the displacement is neither parallel to the strike nor to the dip, and, in such cases the fault is classified as an **oblique-slip fault** (Figure4- 6).

There are two types of strike-slip and two types of dip-slip fault. The two types of strike-slip fault are **right-lateral** (or **dextral**) and **left-lateral** (or **sinistral**) (Fig 4-5).

Strike-slip faults are vertical (or nearly vertical) fractures where the blocks have mostly moved horizontally. If the block opposite an observer looking across the fault moves to the right, the slip style is termed right lateral; if the block moves to the left, the motion is termed left lateral. While the two types of dip-slip fault are **normal** and **reverse** (or **thrust**) . In general, strike-slip faults tend to have dips that are near vertical while dip-slip faults tend to dip about 60° for normal and 30° for reverse or thrust faults. For dip-slip faults, the block lying on top of the fault surface is referred to as the **hanging wall** while the one below is referred to as the **footwall** block.

The first step in classifying dip-slip faults is to identify the hanging and footwall blocks, completing this task, simply note that in normal faults the hanging wall moves down relative to the footwall while for reverse or thrust faults the hanging wall moves up .

In order to classify a strike-slip fault you must first identify some linear feature that is transected by the fault. The location where this feature *pierces* the fault surface is called the **piercing point**. A piercing point is defined as a feature (usually a geologic feature, preferably a linear feature) that is cut by a fault, then moved apart.

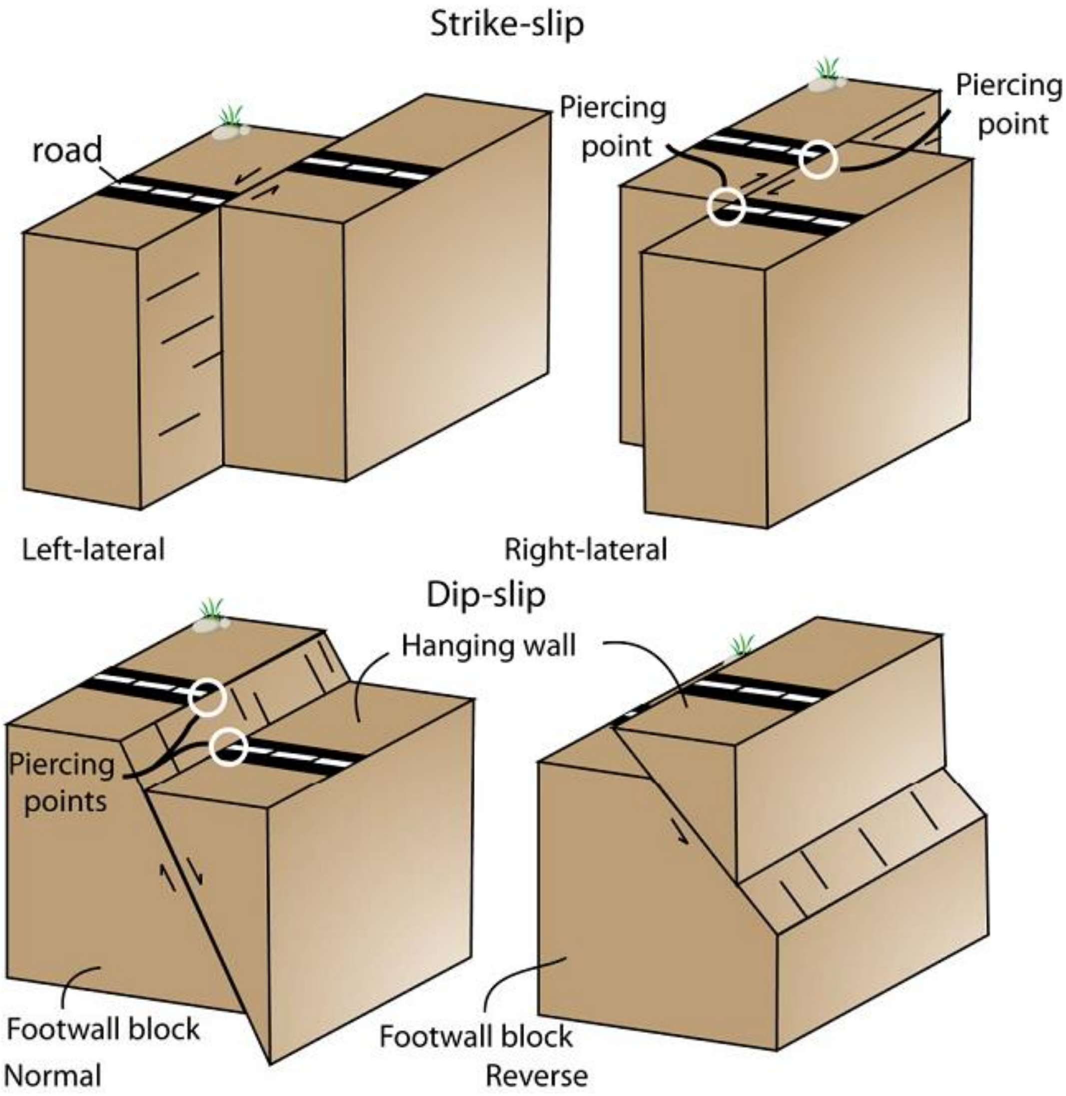


Fig. (4-5)

Oblique-slip faults are classified on the basis of which of the two major components, strike- or dip-slip dominate (Figure 6). If the strike component dominates, then the fault would be classified as either an **oblique right-lateral** or an **oblique left-lateral** strike-slip fault depending upon if the slip is dextral or sinistral respectively (Fig.4-6).

If on the other hand, the dip component dominates, then the fault would be classified as either an **oblique normal** or an **oblique reverse** fault depending upon if the hanging wall moved down or up relative to the footwall block respectively.

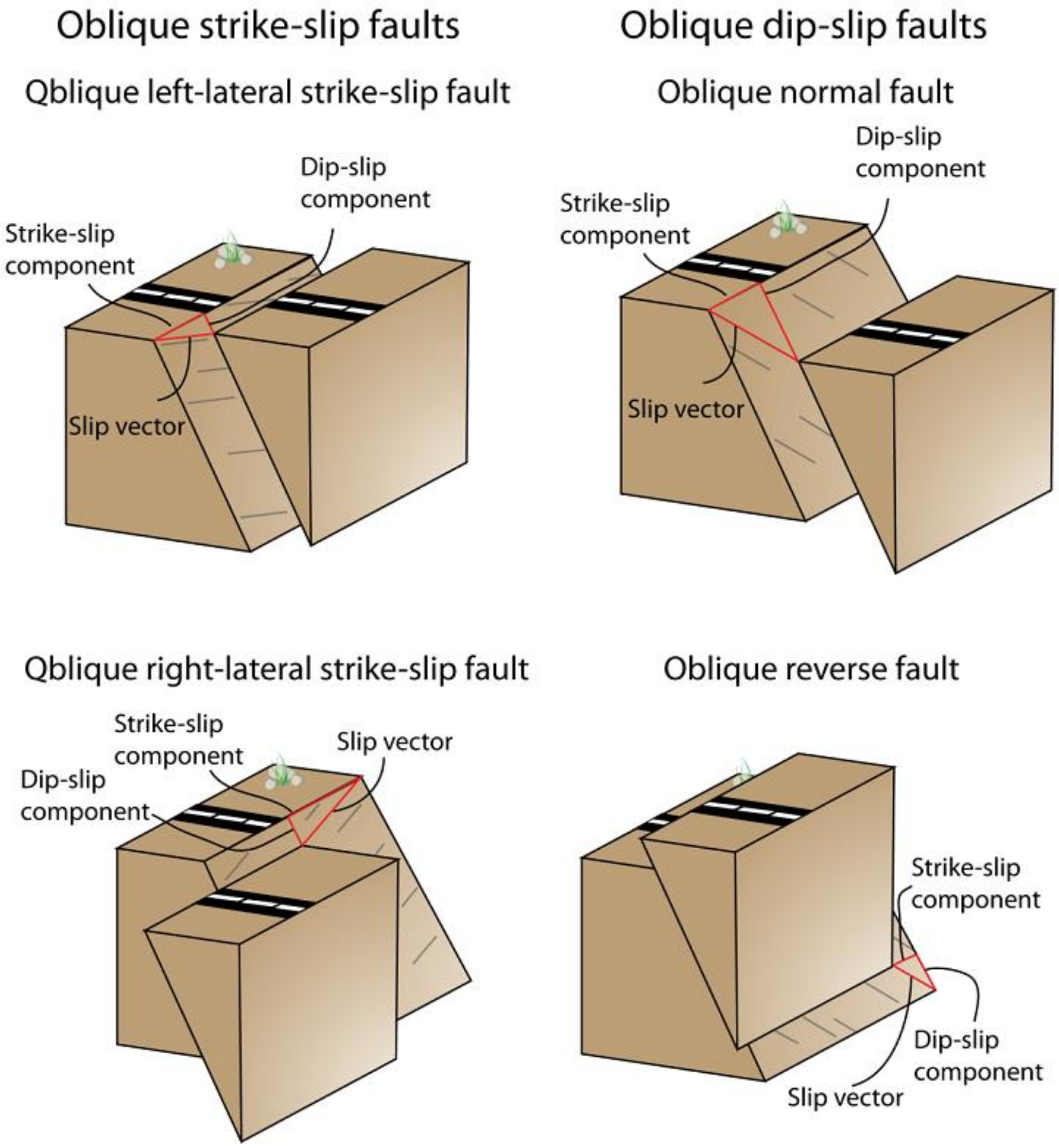


Fig. (4-6)

Normal Fault

Normal Fault: The hanging wall has moved down relative to the foot wall **(Fig 4-7).**

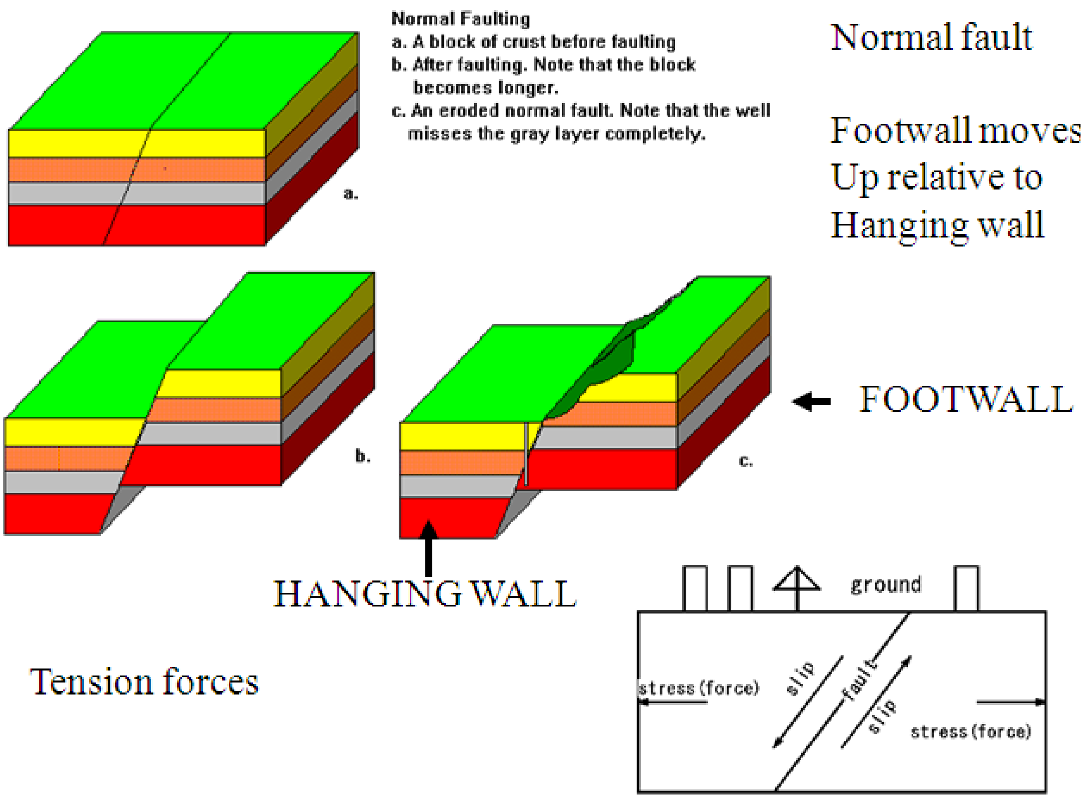
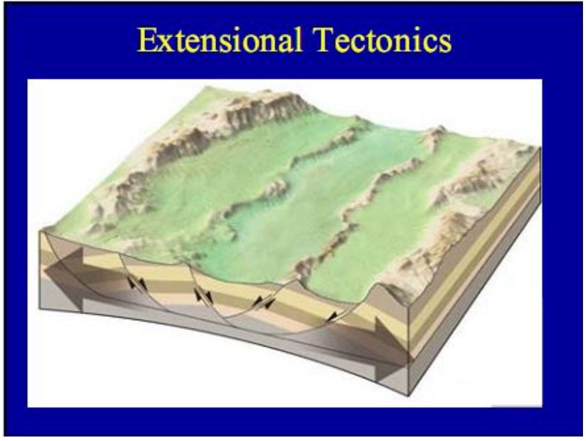


Fig. (4-7)

Tectonic Settings for Extension

- Divergent plate motions
- Gravitational collapse
  - Over-thickened crust
  - Continental margins
- Salt Domes



Extensional Tectonics

Fig.(4-8)shows continental graben formed by normal faults.

A downthrown block between two normal faults dipping towards each other is called a **graben** (Fig.4-8). An upthrown block between two normal faults dipping away from each other is called a **horst** (Fig 4-9, 4-10, 4-11, 2-12). Low-angle normal faults with regional tectonic significance may be designated **detachment faults** (Fig 4-13).

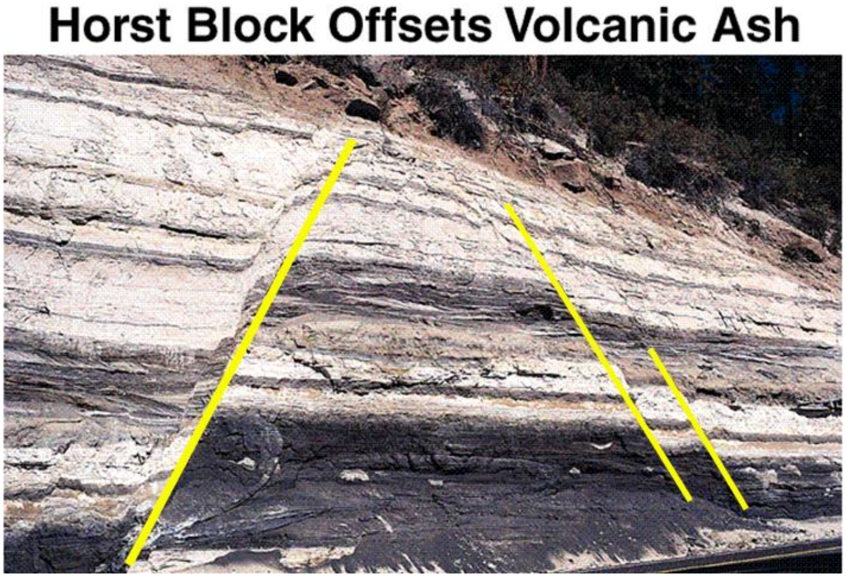


Fig.(4-9) Shows horst formed by normal faults

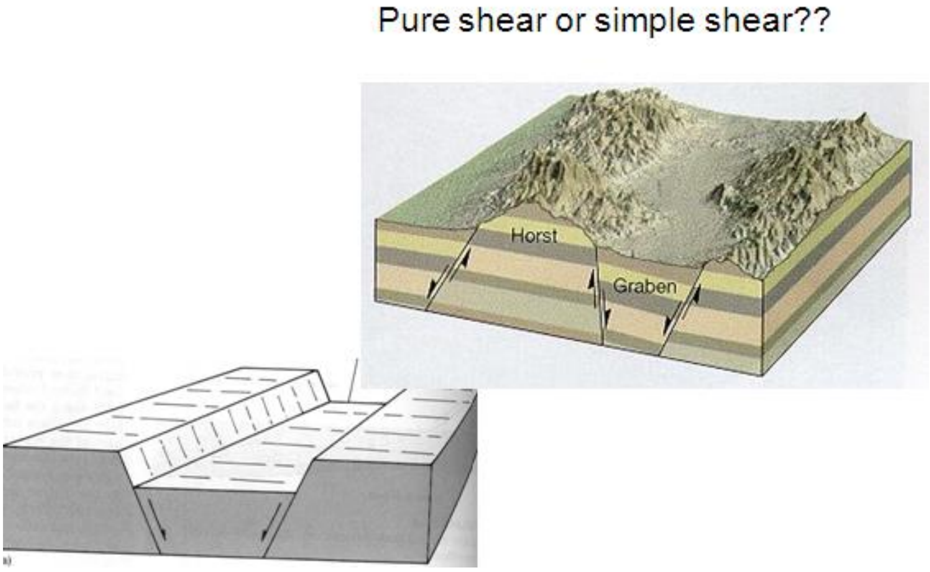
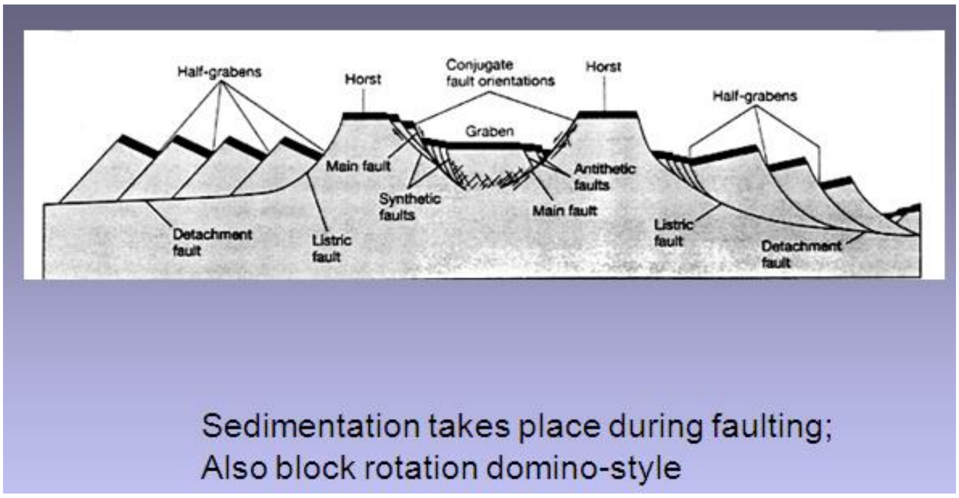


Fig. 4-10 Shows horst and graben.

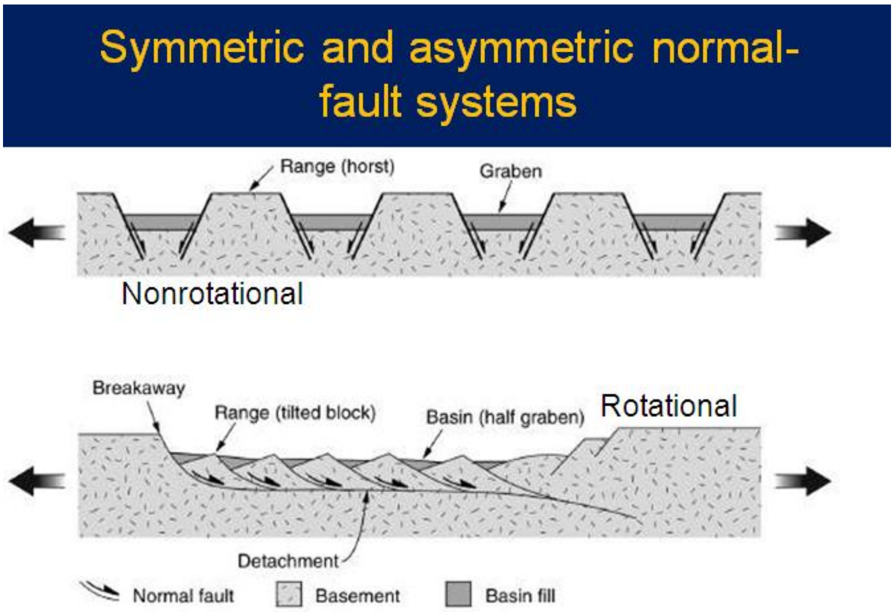
Fig. (4.11) Shows photo of horst.



Sedimentation takes place during faulting;
Also block rotation domino-style

Fig.(4.12) shows graben , horsts and half grabens in section.

Fig. 4-13 Shows photo of detachment fault.



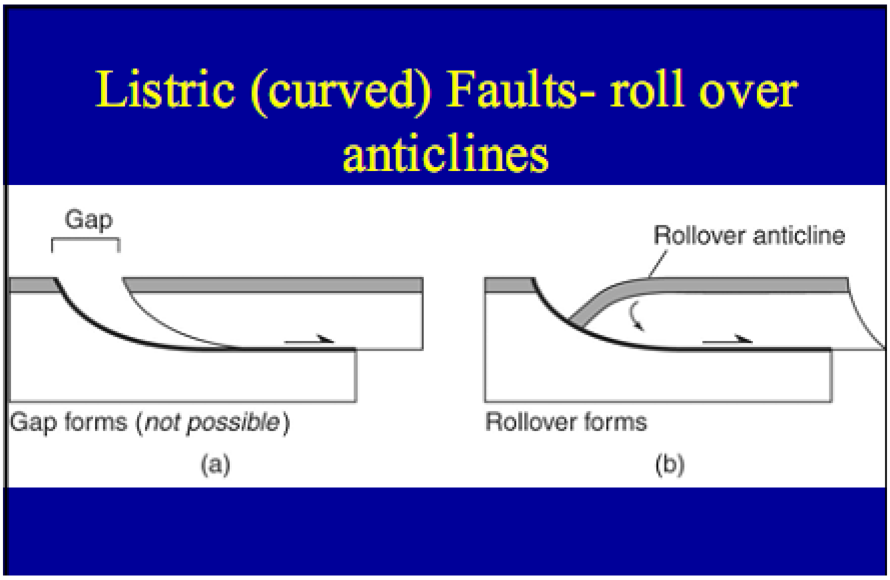
Symmetric and asymmetric normal-fault systems

Fig. (4-14) shows symmetric and asymmetric normal fault systems.

Listric Faults

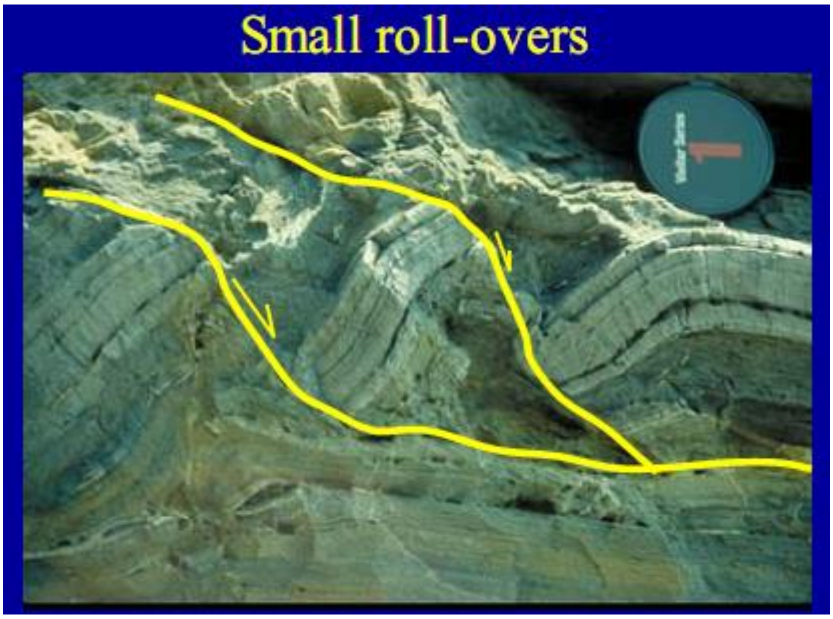
The second main type of faults found in extensional regimes, listric faults can be defined as curved normal faults in which the fault surface in concave upwards; its dip decreases with depth. These faults also occur in extension zones where there is a main **detachment** fracture following a curved path rather than a planar path (Fig. 4-14). Hanging wall blocks may either rotate or slide along the fault plane (e.g. slumps), or they may pull away from the main fault, slipping instead only along the low dipping part of the fault. **Roll-over anticlines** will often form between bedding planes and the main fault plane as a result of the flexing between the two (Fig. 4-15).

A **listric** fault is a type of fault in which the fault plane is curved. The dip of the fault plane becomes shallower with increased depth and may flatten into a sub-horizontal décollement.



Listric (curved) Faults- roll over anticlines

Fig. (4-14)



Small roll-overs

Fig.(4-.15)Shows roll-overs anticlines, formed by listric faults.

Reverse Faults

A **reverse fault** is the opposite of a normal fault, the hanging wall moves up relative to the footwall. Reverse faults indicate compressive shortening of the crust. The dip of a reverse fault is relatively steep, greater than 45° Figs (4-16, 17A).

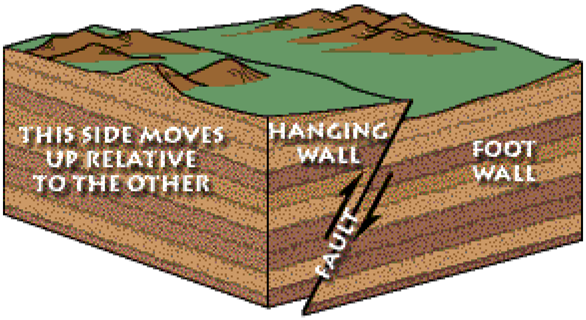


Fig. (4-16), shows reverse fault.

A **thrust fault** has the same sense of motion as a reverse fault, but with the dip of the fault plane at less than 45° Fig. (4-17C). Thrust faults typically form **ramps**, **flats** and **fault-bend** (hanging wall and foot wall) folds. Thrust faults form nappes and klippen in the large thrust belts (Figs. 4-19, 4-20).

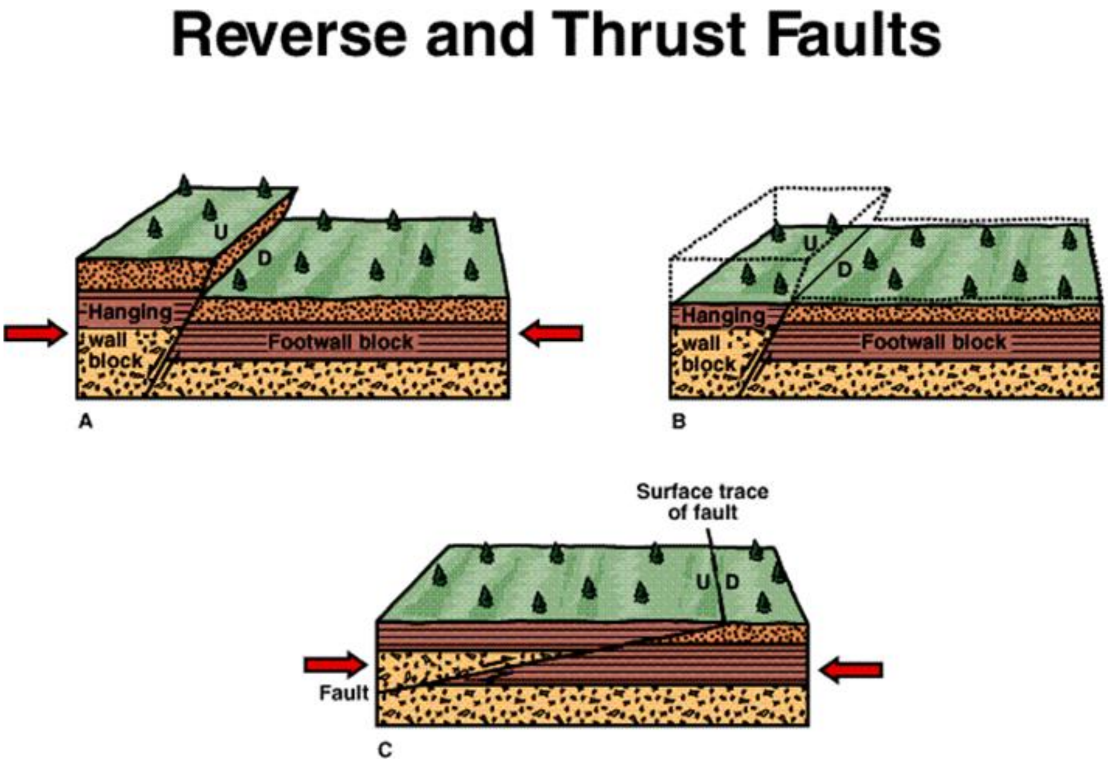
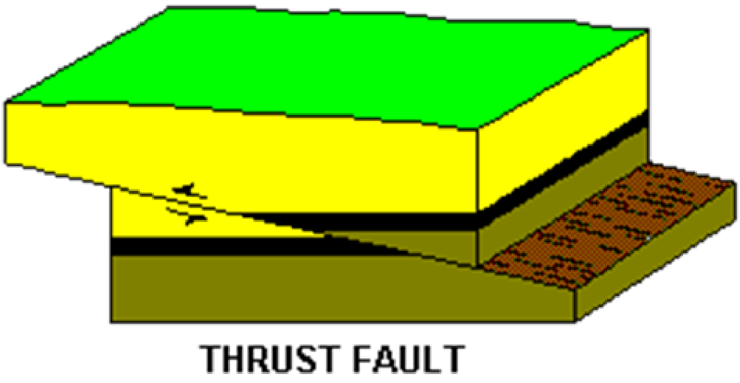

Reverse and Thrust Faults

Fig.(4-17)Shows(A) reverse fault, (B) reverse fault affected by erosion, (C) thrust fault

Subduction zones are a special class of thrusts that form the largest faults on Earth and give rise to the largest earthquakes (Fig. 4-18).

REVERSE FAULTS: Hanging wall moves up relative to footwall
Result of compression: plates colliding
Two types: low-angle or thrust faults, and high-angle reverse faults



Individual layers can move 100's of kilometers
Alps are a great example

Fig.(4-18) Shows thrust fault

The fault plane is the plane that represents the fracture surface of a fault. Flat segments of thrust fault planes are known as ***flats***, and inclined sections of the thrust are known as ***ramps***. Typically, thrust faults move *within* formations by forming flats, and climb up section with ramps (Fig. 4-19).

Fault-bend folds: Thrust faults, particularly those involved in thin skinned style of deformation, have a so-called **ramp-flat geometry**, are formed by movement of the hanging wall over a non-planar fault surface and are found associated with both extensional and thrust faults. Thrusts mostly propagate along zones of weakness within a sedimentary sequence, such as mudstones or salt layers; these parts of the thrust are called **flats**. If the effectiveness of the **decollement** becomes reduced, the thrust will tend to cut up the section to a higher stratigraphic level until it reaches another effective decollement where it can continue as bedding parallel flat. The part of the thrust linking the two flats is known as a **ramp** and typically forms at an angle of about 15°-30° to the bedding. Continued displacement on a thrust over a ramp produces a characteristic fold geometry known as a **ramp anticline** or, more generally, as a **fault-bend fold**, (Fig 4-19).

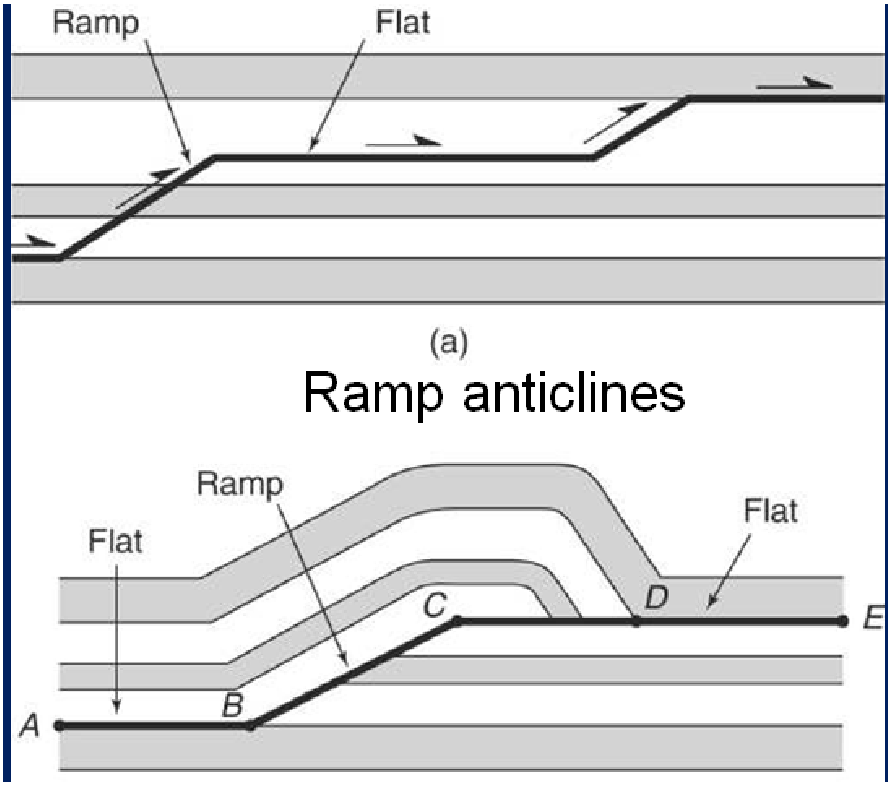


Fig. 4-19 Shows the Flats and Ramp in thrust fault.

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

Décollement (/dɛ.kɒll.mɔːn/; from the French décoller, 'to detach from') is a gliding plane between two rock masses, also known as a basal detachment fault. Décollements are a deformational structure, resulting in independent styles of deformation in the rocks above and below the fault. They are associated with both compressional settings (involving folding and overthrusting) and extensional settings.

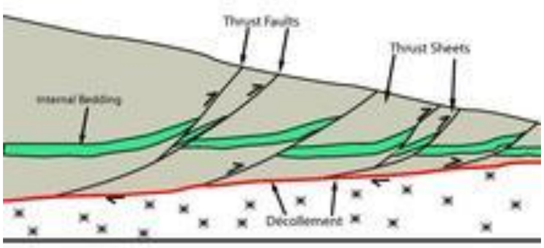


Fig.(4-20)Shows Decollement structure.

Fault-propagation folds

Fault-propagation folds form at the tip of a thrust fault where propagation along the decollement has ceased but displacement on the thrust behind the fault tip is continuing. The continuing displacement is accommodated by formation of an asymmetric anticline-syncline fold pair. As displacement continues the thrust tip starts to propagate along the axis of the syncline. Such structures are also known as **tip-line folds**. Eventually the propagating thrust tip may reach another effective decollement layer and a composite fold structure will develop with characteristics of both **fault-bend** and **fault-propagation folds Fig,(4-21)**.

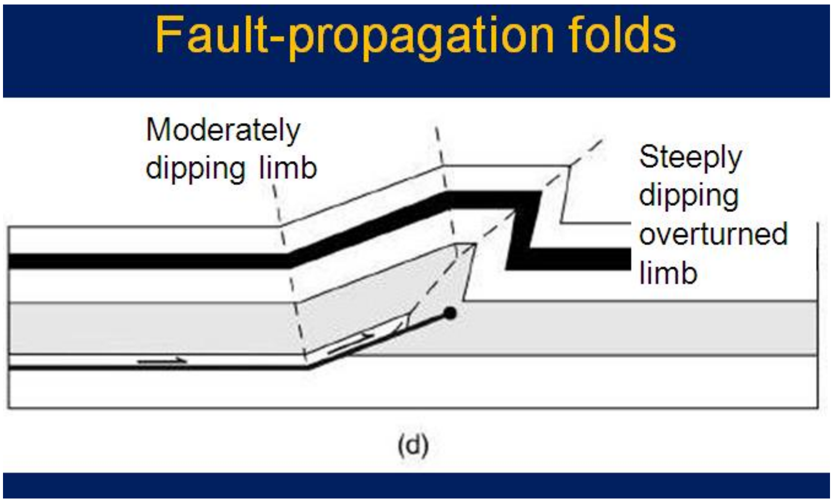
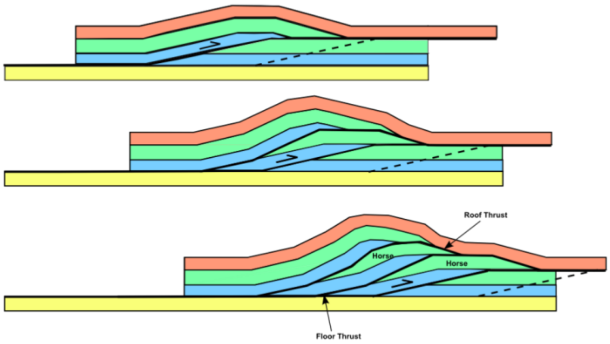


Fig.(4-21)Shows Fault-propagation fold.

Thrust duplex

Duplexes occur where there are two decollement levels close to each other within a sedimentary sequence, such as the top and base of a relatively strong sandstone layer bounded by two relatively weak mudstone layers. When a thrust that has propagated along the lower detachment, known as the **floor thrust**, cuts up to the upper detachment, known as the **roof thrust**, it forms a ramp within the stronger layer. With continued displacement on the thrust, higher stresses are developed in the footwall of the ramp due to the bend on the fault Fig,(4-22).



Fig,(4-22), Shows Thrust duplex.

Blind thrust faults

If the fault plane terminates before it reaches the Earth's surface, it is referred to as a blind thrust fault. Because of the lack of surface evidence, blind thrust faults are difficult to detect until they rupture Fig.(4-23).

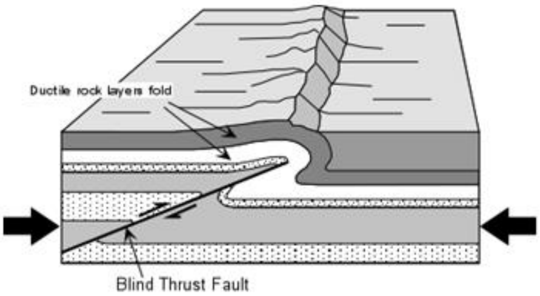
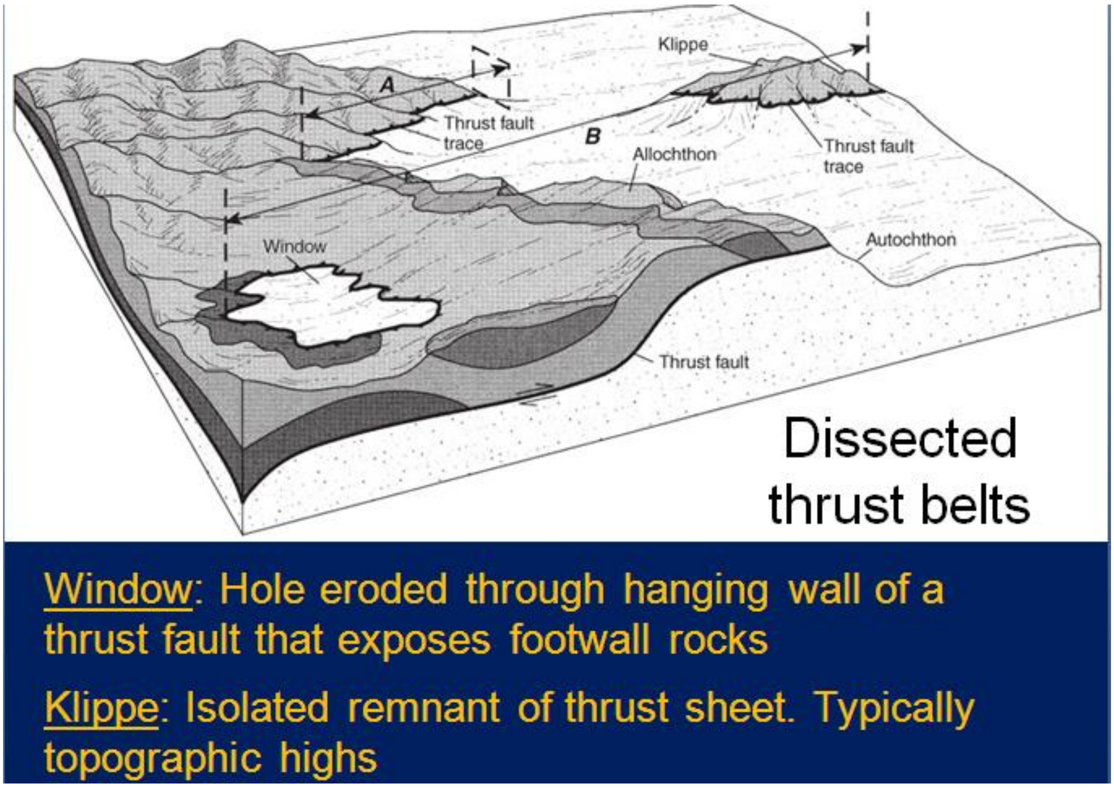


Fig.(4-23)Shows Blind Thrust fault.

Klippe and Window

A **klippe** (German for cliff or crag) is a geological feature of thrust fault terrains. The klippe is the remnant portion of a nappe after erosion has removed connecting portions of the nappe. This process results in an outlier of exotic, often nearly horizontally translated strata overlying autochthonous strata. A fault **inlier**, **fenster**, or **window** is an area of the **autochthonous** basement uncovered by erosion, but continuously surrounded by the body of the nappe; the Hohe Tauern window in the Alps is a typical example.

Allochthon, or an allochthonous block, is a large block of rock which has been moved from its original site of formation, usually by low angle thrust faulting. An allochthon which is isolated from the rock that pushed it into position is called a klippe. If an allochthon has a "hole" in it so that one can view the autochthon beneath the allochthon, the hole is called a "window" (or Fenster) Figs,(4-24, 4-25).



Window: Hole eroded through hanging wall of a thrust fault that exposes footwall rocks

Klippe: Isolated remnant of thrust sheet. Typically topographic highs

Fig.(4-24) Shows Window and Klippe landforms resulted from thrust fault.

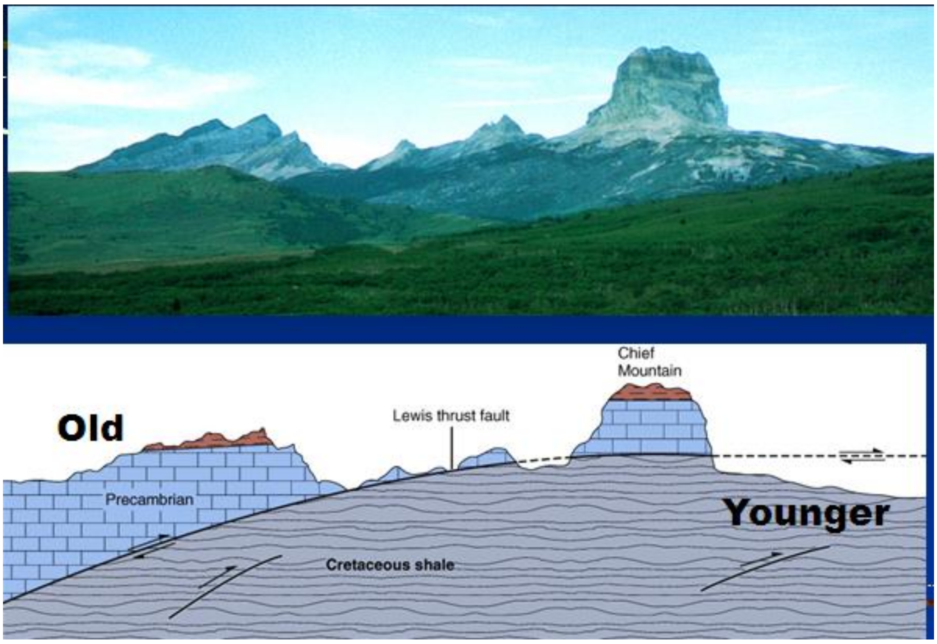
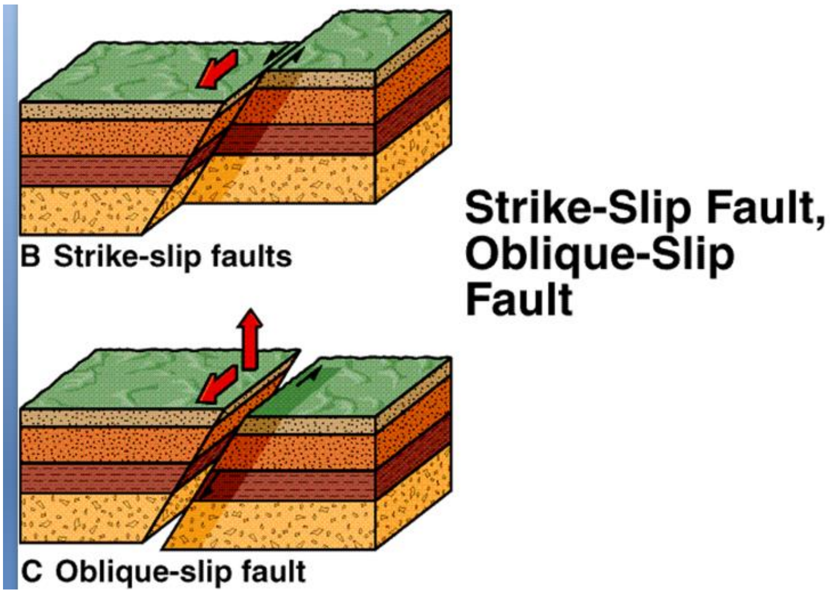


Fig.(4-25) Shows photo and sketch of Window and Klippe.

Strike-Slip Fault



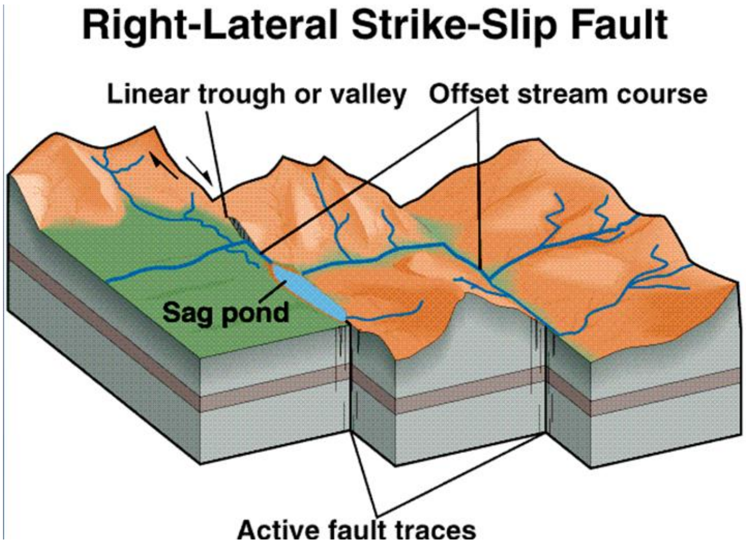


Fig.(5.1) Shows oblique left lateral strike slip faults and en echelon right lateral active faults.

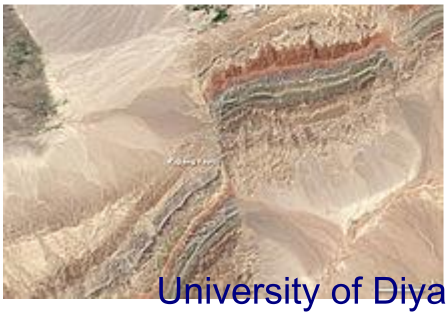


Fig.(5.2) shows photo of right lateral strike slip fault

Linkage

Strike slip faults are commonly segmented at all scales and levels of exposure, typically in the form of *en échelon*, non-coplanar faults separated by **offsets** (or **step-overs**). These step-over zones of host rock between the end and the beginning of two adjoining *en échelon* shear fractures deform in order to accommodate continued strike slip displacement. This local deformation may lead to the formation of short fault segments that connect adjacent *en échelon* fault segments and result in a through-going fault zone. The geometry of these step-over zones and linking faults, in turn, controls **contractional** or **extensional** deformation according to:**1- the sense of slip** and **2-stepping direction of the *en échelon* fault segments**.

Left-stepping refers to the arrangement in which one fault segment occurs to the left of the adjacent segment from which it is being viewed. The contrary is **right-stepping (Fig.5.3)**. **Contractional** or **restraining bends** and offsets are local zones of convergence where material is pushed together by the dominant fault movement. The linkage of adjacent fault segments is typically through the development of **P-shear** splay faults. At a constant volume of the deforming transpression zone, local shortening will produce vertical lengthening and thus surface uplift. This **push-up** area will be eroded.

- **Extensional**, **releasing** or dilatant bends and offsets are local zones of extension where material is pulled apart by the dominant fault movement. The linkage of adjacent fault segments is typically through the development of **R-shear** splay faults. At a constant volume of the deforming transtension zone, local extension will produce vertical shortening and surface depression. This **pull-apart** area will be site for sedimentation (Fig.5.3).

A strike-slip fault system commonly shows a braided pattern of **anastomosing** contemporaneous faults. Contractional and extensional bends and offsets can thus alternate along a single yet complex strike-slip zone.

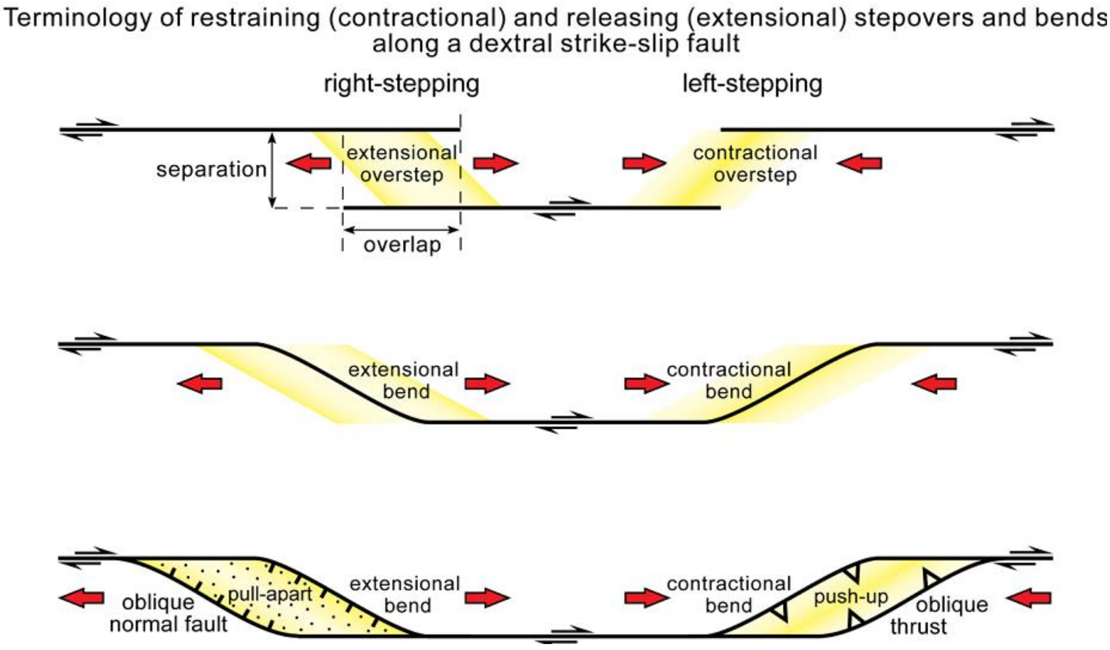


Fig.(5.3)

Subsidiary fractures = Riedel shears

Subsidiary shear fractures that propagate a short distance out of the main fault but are coeval with it are called Riedel shears. This term is also used on a large-scale fault pattern and may refer to as many as **five direction** families of associated fractures Fig 5-3A. In that case, individual fractures remain active after the other types developed so that synchronous movement on all fractures accommodate strain in the fault zone. The geometrical arrangement of Riedel shears is indicative of the sense of movement within the wrench zone and is therefore widely used for the interpretation of its kinematic evolution.

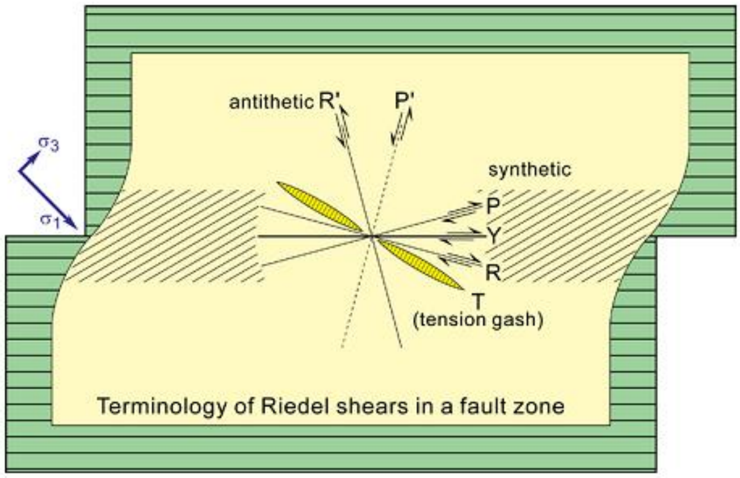


Fig. 5-3A Shows Riedel Fractures.

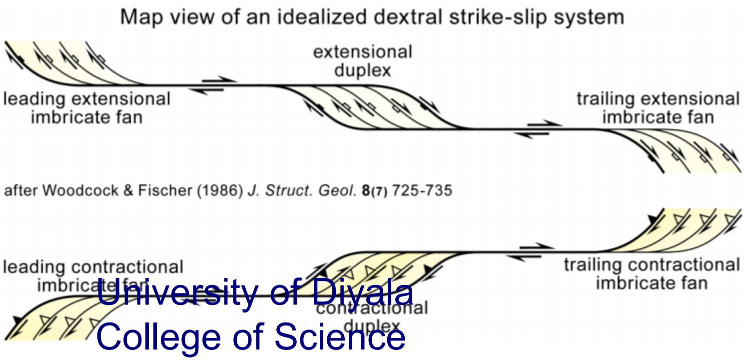
- **R Riedel shears** are generally the first subsidiary fractures to occur and generally build the most prominent set. They develop at an acute angle, typically 10-20° clockwise to a dextral main fault, anticlockwise to a sinistral strike-slip fault. They often form an en échelon and overstepping array synthetic to the main fault; they evolve as a sequence of linked displacement surfaces. Their acute angle with the fault points in the direction of the relative sense of movement on the main fault. This angle is equal to $\varphi/2$, where φ is the material internal friction angle.

R' shears are antithetic faults (i.e. with a sense of displacement opposite to the bulk - movement) oriented at a high angle (approximately 75°, i.e. $90^\circ - \varphi/2$ clockwise to a dextral, anticlockwise to a sinistral main fault plane), conjugate with the R(Riedel) shears. They preferentially occur in the overlap zone between two parallel R shears and often connect these two R shears. They may develop **with** or **after** R shears.

P shears are synthetic minor faults symmetrically oriented to the R shears with - respect to the fault plane (at $\varphi/2$ from the fault plane, anticlockwise and clockwise to dextral and sinistral faults, respectively). P shears also form an en échelon array contemporaneous with R shears or later as links between R shears. P-shears are contractional and accommodate fault parallel shortening as shearing proceeds. They are less common as R and R' shears and may require more displacement to form. As for R Riedel shears, there may be **P' shears** conjugate with P shears but these have relative minor importance and are difficult to separate, in terms of orientation, from R-shears.

- **Y shears** are synthetic micro faults sub parallel to the main fault, apparently the **last** to form. Riedel micro faults may all connect one another to form an anastomosing network of fractures in a narrow fault zone whose bulk borders are parallel to the main fault.

Strike-slip duplexes*:* Multiple linking of closely-spaced **R- and P-shears** may create fault-bound lenses (elongate horses) imbricated between overlapping en échelon segments. Such sets of horizontally stacked and isolated rock lenses are bounded on both sides by parallel segments of the main fault and thus define **strike-slip duplexes** (like thrust or normal-fault duplexes, but tilted to the vertical). They develop in **transfer zones**, where displacement is conveyed from one fault segment to another in systems of stepped strike-slip faults, and in **bends**, where the orientation of the main fault is deflected (Fig.5.4).



after Woodcock & Fischer (1986) *J. Struct. Geol.* **8(7)** 725-735

Fig.(5.4)

Strike-slip duplexes may be **compressional** or **extensional**, depending on whether they formed at an extensional (facing towards the movement direction) or contractional (facing against the relative movement) bend. **Thrust** or **normal**-fault duplexes accommodate vertical **thickening** (through stacking of vertical slabs that rise upward and outward over the adjacent blocks) or **thinning** (through separation of horses) of the crust. For strike-slip duplexes, the corresponding thickening or thinning would have to occur in a horizontal direction, which is difficult owing to the constraint imposed by the rest of the crust. The required deformation can be easier accommodated vertically, and, therefore, strike-slip duplexes involve oblique movements. *In a compressional strike-slip duplex, fault must combine strike- and reverse slip; in an extensional strike-slip duplex, faults combine strike- and normal slip (Fig.5.5).*

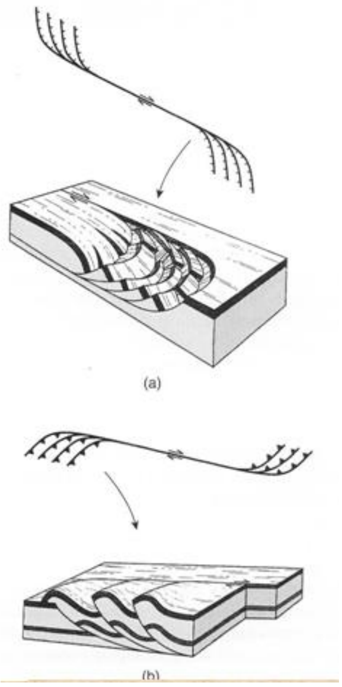


Fig.(5.5)

Scissor-faults

Rotation of horses around a horizontal axis may produce **scissor-faults**, which change from a normal fault at one end to a reverse fault at the other (Fig.5.6). Duplexes are commonly breached by faults that connect the stepping segments.

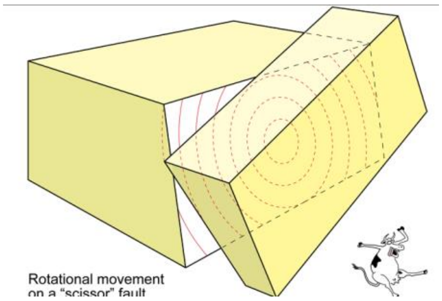


Fig.(5.6)

Horsetail splay: Like any other fault, strike-slip faults may terminate in zones of **ductile deformation**.

In brittle terminations, the displacement is distributed through several branching splay faults. These small faults, curved away from the strike of the main fault, form an open, imbricate fan called a **horsetail splay**. Antithetic and synthetic splay faults at tips of major strike slip faults have often a small vertical component consistent with the extensional or compressional character of the fault termination. Large-scale, extensional horsetail splays may host **sedimentary basins** at tips of major strike-slip faults. Conversely, compressional horsetail splays may display **thrust faults** and **folds** at tips of major strike-slip faults (Fig5.7).

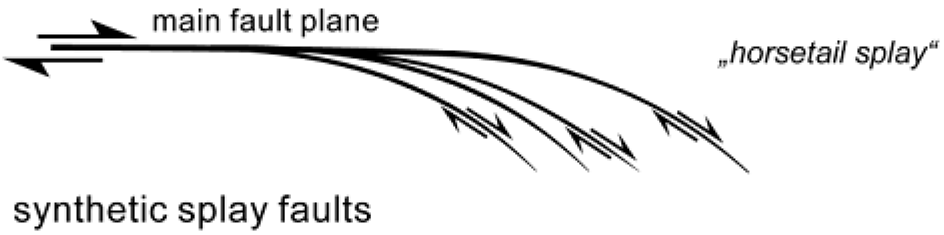


Fig.(5.7)

Flower structures

Seismic profiles across main faults of transpressive and transtensive strike-slip duplexes have revealed the following characteristics: - Fan-like, rather steep faults converge at depth into a single and subvertical fault. - The deep main fault (the **stem**) is subvertical. Normal and reverse offsets along a single fault plane often result from inversion of the relative movement on the fault. This upward splay shape of subsidiary faults is termed a **flower structure**. - If the vertical component is **normal,** faults tend to be listric and to form a normal or **negative flower structure**, which forms a depressed area. It forms a **sag pond,** a **rhomb graben** or, on a larger scale, a **pull-apart** basin. Negative flower structures are also called **tulip structures** (Figs. 5.8, 5.9, 5.10, 5.11).

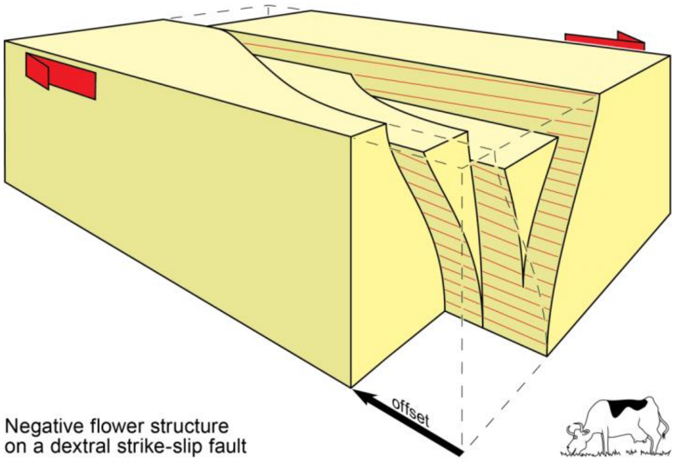
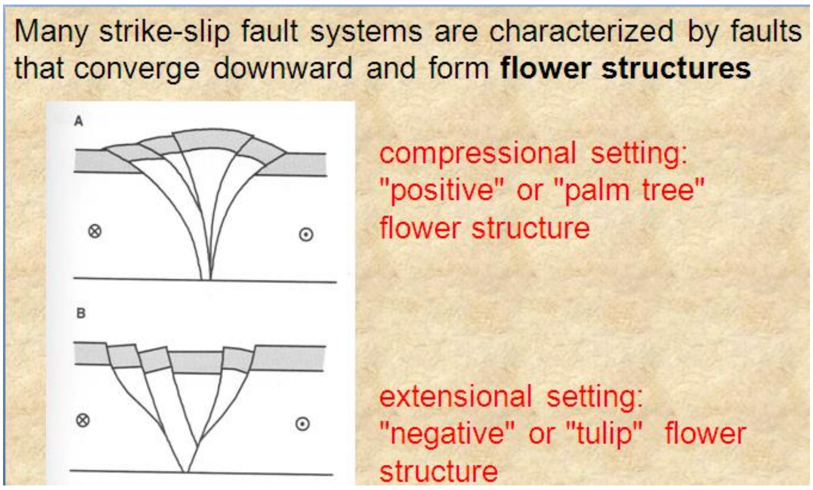


Fig.(5.8)

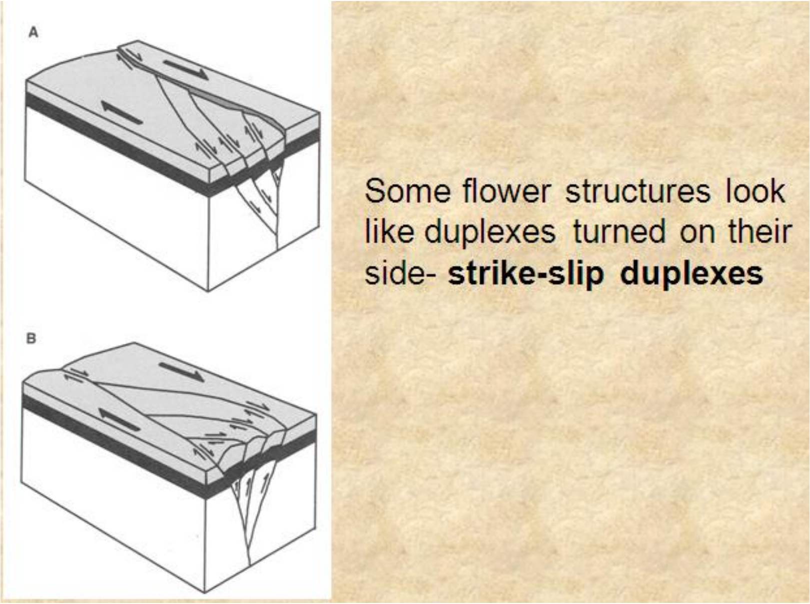


Many strike-slip fault systems are characterized by faults that converge downward and form **flower structures**

compressional setting: "positive" or "palm tree" flower structure

extensional setting: "negative" or "tulip" flower structure

Fig.(5.9)



Some flower structures look like duplexes turned on their side- **strike-slip duplexes**

Fig.(5.10)

If the vertical component is reverse, the splay faults tend to be convex upward, with at the surface. They form a reverse or **positive flower structure**, which appears as an uplifted, commonly antiformal area (a **rhomb horst** or **push-up**). Positive flower structures are also termed **palm-tree structures**, owing to the convex upward form of the upward-diverging faults.

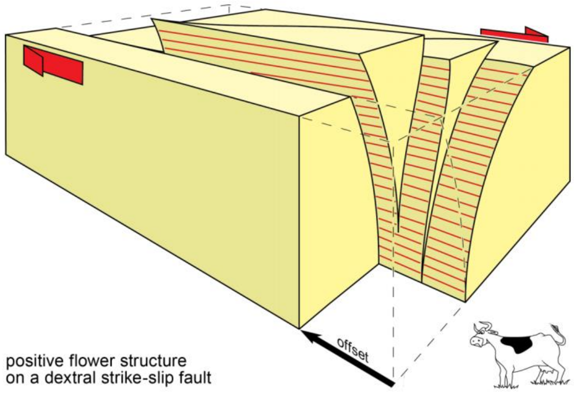


Fig. (5.11)

Relationship between folds and strike-slip faults *Passive en échelon folds*

Folds associated to **wrench** fault systems are typically non-cylindrical, doubly plunging and relatively short with steeply dipping axial planes. They are arranged spatially such that *culminations* and *depressions* in successive folds lie along lines that make an acute angle with the approximately parallel fold axes. Such folds are stepped, consistently overlapping, and said to be arranged *en échelon*. Taking the axial planes as roughly orthogonal to the shortening direction, their distribution permits to decipher the potential strike-slip fault they are related to. **Such folds are common above strike slip faults in the basement**, **which have not broken the cover**. The *en échelon* folding reveals the relative sense of movement (Fig.5.12).

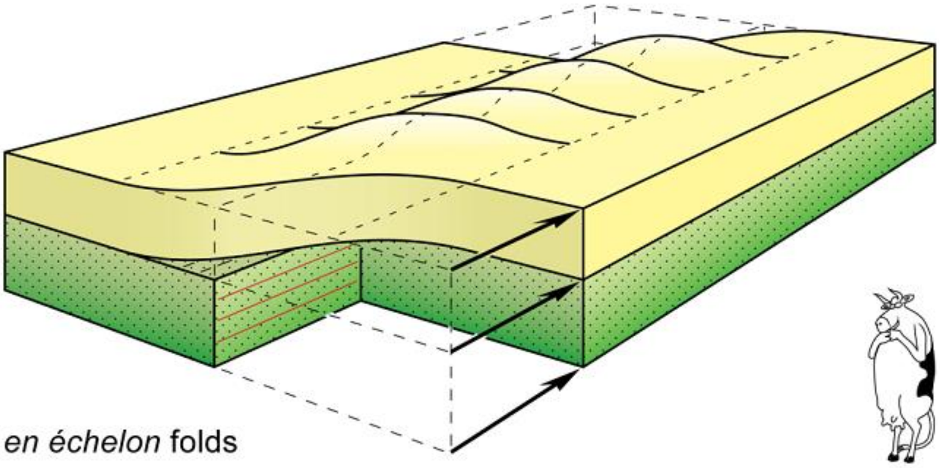
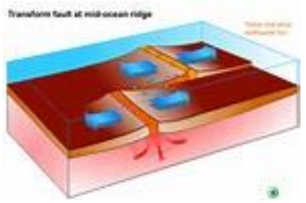
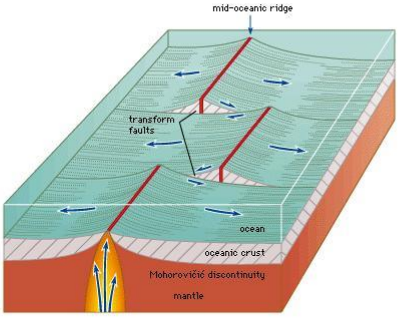


Fig.(5.12)

Transform Faults: Are a type of strike-slip fault (defined by Wilson 1965). They are due to the differences in motion between lithospheric plates. They are named transform because they are first noticed in oceans linking areas where crust is transformed into mantle, or mantle is transformed into crust or the opposite or are themselves plate boundaries. They are basically occur where type of plate boundary is transformed into another. The main types of transform faults are:

1) Ridge-Ridge 2) Ridge-Arc 3) Arc-Arc





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Transfer fault

A **transfer fault** is a strike slip fault that transfers displacement between two similar non-coplanar structures (e.g. in step-overs between two normal or thrust faults). It is striking parallel to the regional direction of extension or compression. The transfer fault terminates on these two other structures and is also known as **lateral ramp**. It is a local and passive fracture formed in response to active faulting on faults which link with the transfer. Hence, transfer faults can reverse their strike slip sense in time and space and may show apparent offsets opposite to the true movement sense (Fig.5.13).

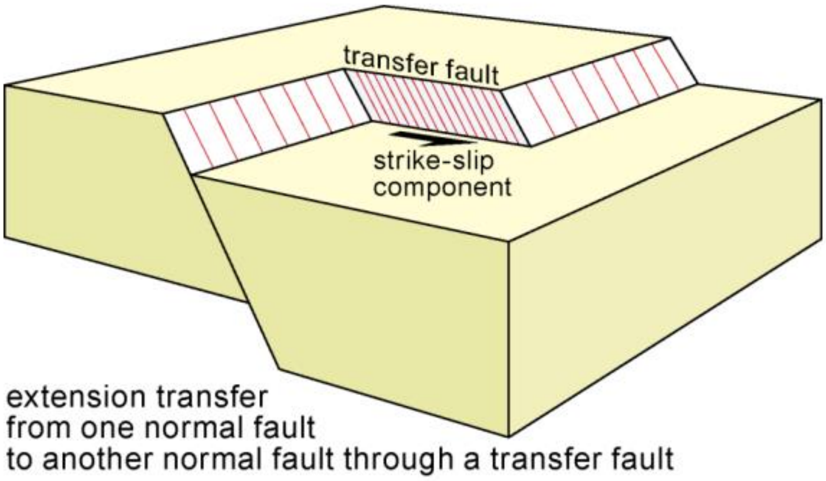


Fig.(5.13)

Tear fault

A **tear fault** is a relatively small strike slip fault that runs across the strike of a contractional or extensional belt and accommodates differential displacement between two adjacent segments of the belt. Tear faults are therefore parallel to the movement direction of thrusts or normal faults; they are usually common in hanging walls of **low-angle faults**. Fold axes, where folding is involved, tend to terminate against tear faults (Fig.5.14).

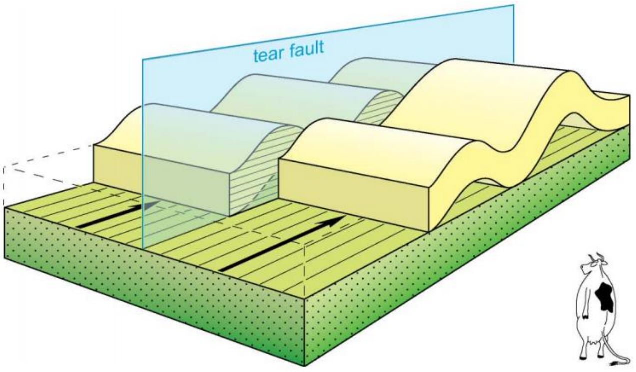
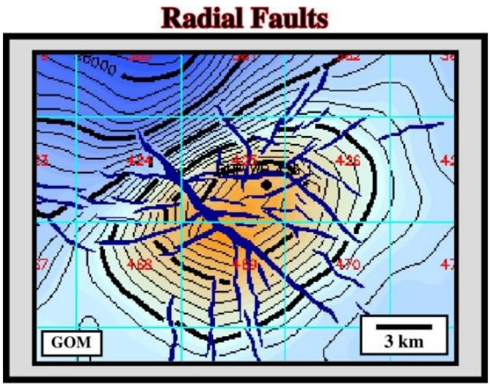


Fig.(5.14)

en-echelon faults: Faults that are approximately parallel one another but occur in short unconnected segments, and sometimes overlapping.

Radial faults: faults that are converge toward one point.



Concentric faults: faults that are concentric to a point.

Bedding faults (bedding plane faults): follow bedding or occur parallel to the orientation of bedding planes.

CRITERIA FOR FAULTING:

1) Repetition or omission of stratigraphic units asymmetrical repetition .

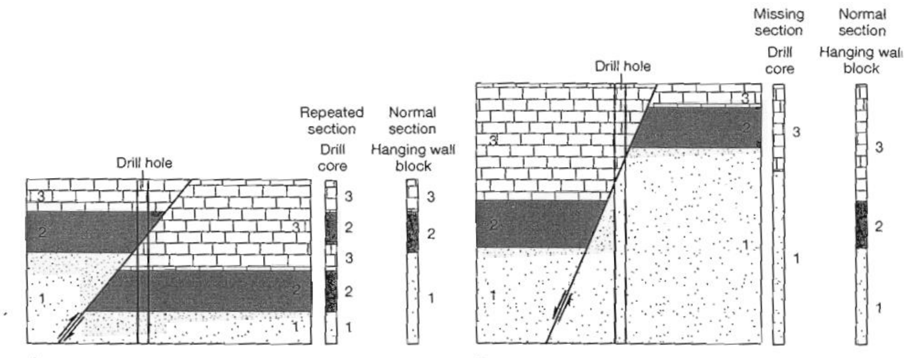


Figure 4.10 *A.* Thrust fault resulting in repeated section in a vertical drill hole. *B.* Normal fault resulting in missing section in a vertical drill hole.

Fig.(6.1)

2) ***Displacement of recognizable marker such as fossils, color, composition, texture..etc..***

Fig.(6.2)

3) ***Truncation of structures, beds or rock units.***

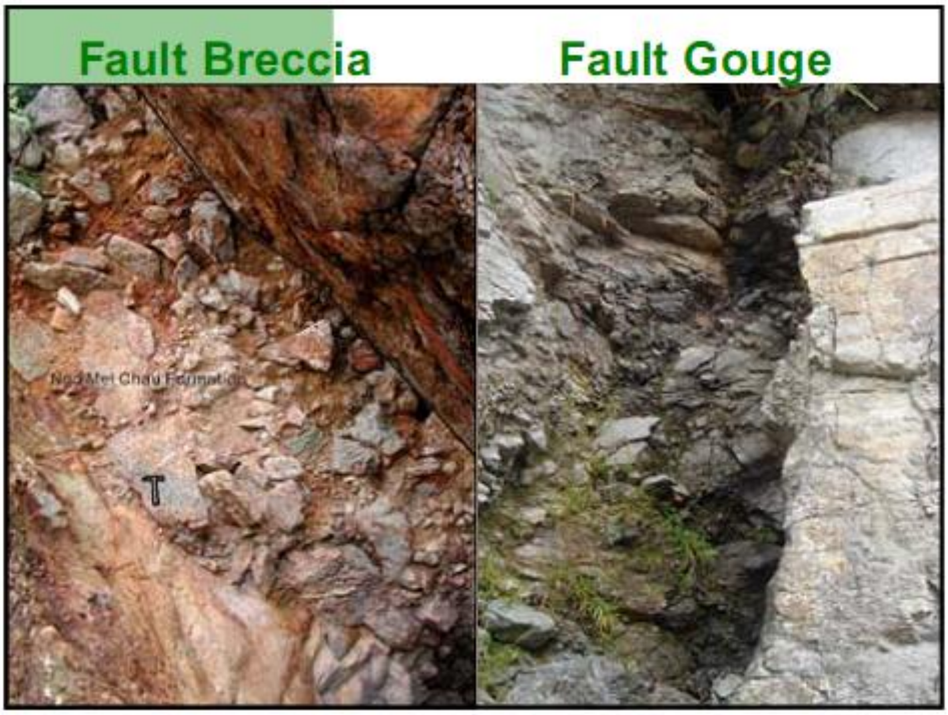


Fig.(6.3)

4) ***Occurrence of fault rocks (mylonite or cataclastic or both)***

5) ***Presence of S or C structures or both, rotated porphyry clasts and other evidence of shear zone.***

6) Abundant veins, silicification or other mineralization along fracture may indicate faulting.

Fig.(6.4)

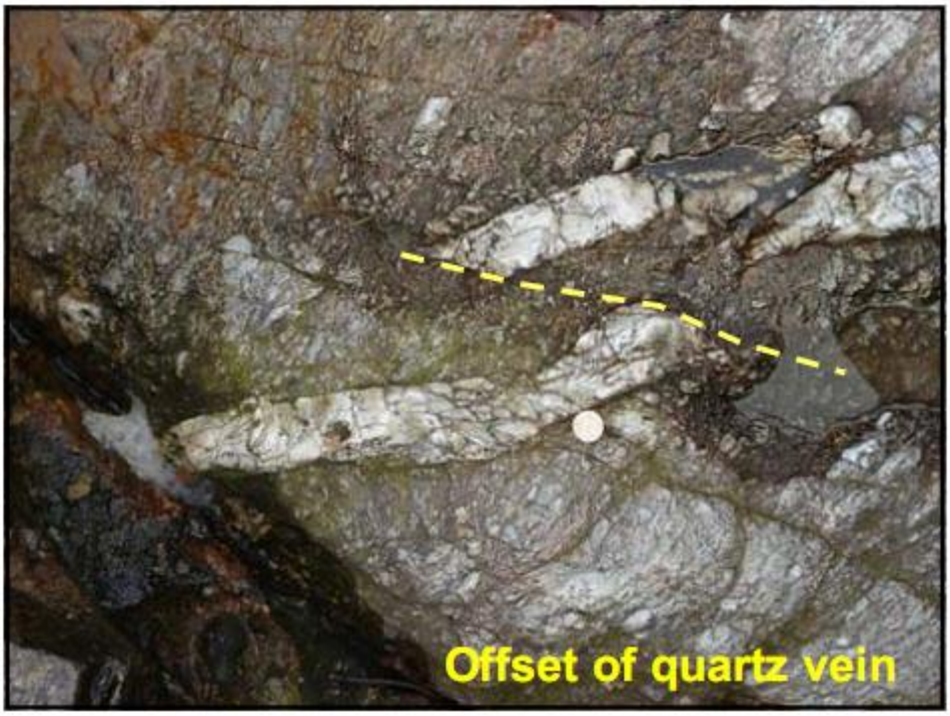
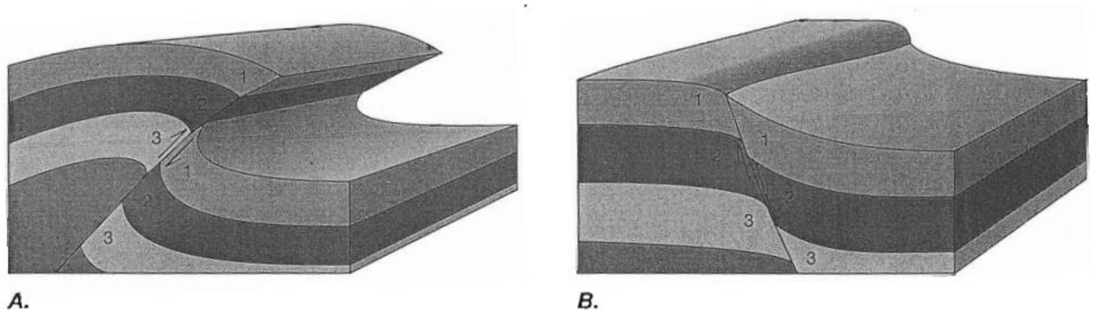


Fig.(6.5)

7) Drag Units appear to be pulled into a fault during movement (usually within the drag fold and the result is thrust fault).



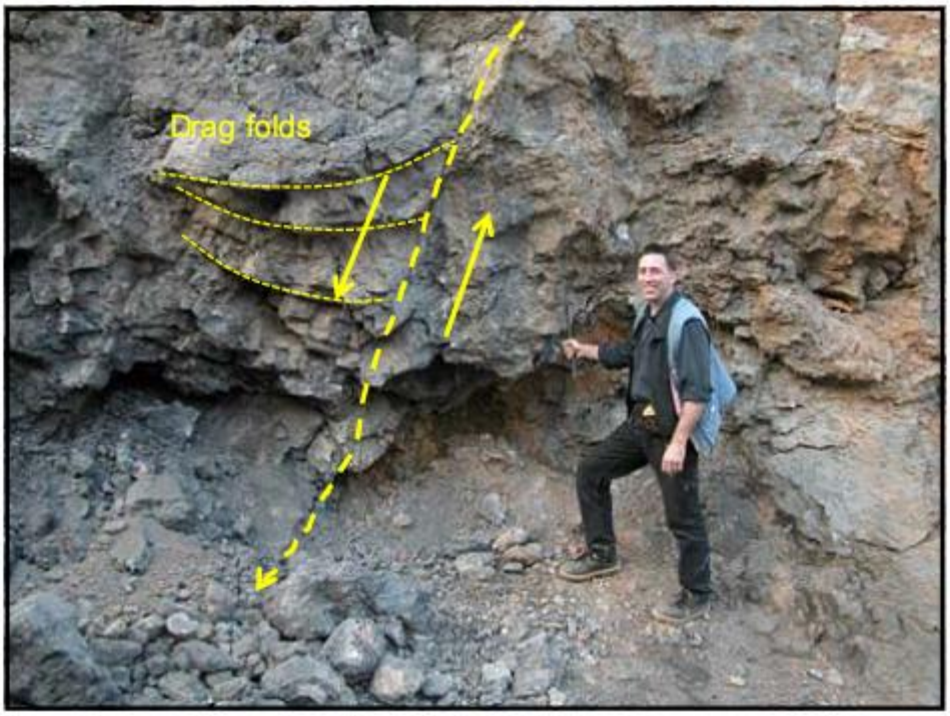


Fig.(6.6)

8) Reverse drag): occurs along listric normal faults.
9) Slickensides and slickenlines along a fault surface

Fig.(6.7)

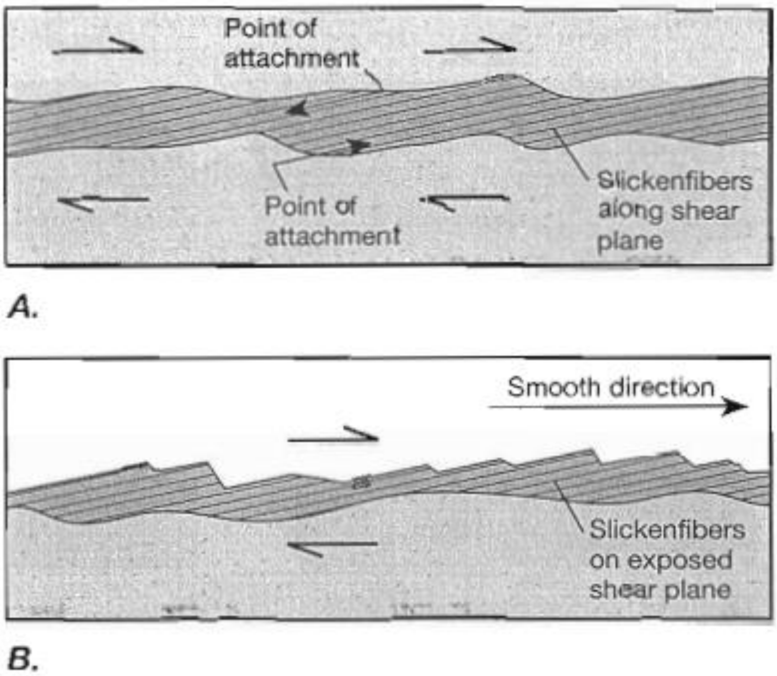
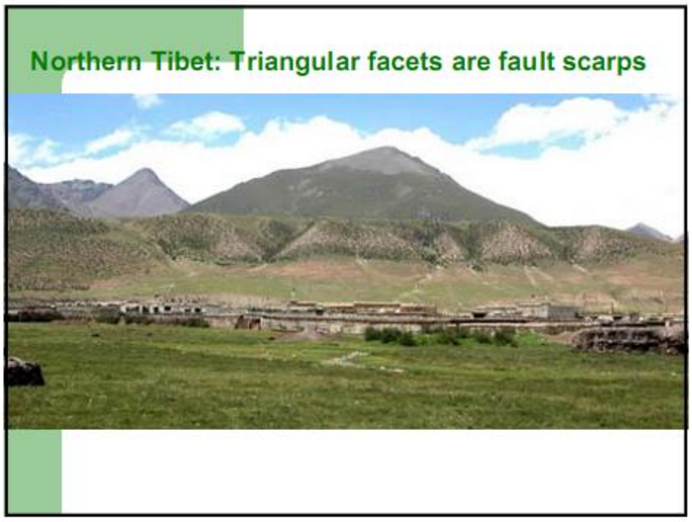


Figure 4.15 Slickenfibers as indicators of shear sense and minimum displacement. *A.* An arrow along a slickenfiber with its base at the point of attachment to one wall of the fault points in the direction of relative slip of the opposite wall of the fracture. The length of the fiber from one wall to the other is a measure of the minimum displacement on that fault. *B.* The smooth, or "downstairs," direction on the stepped surface of an exposed set of slickenfibers defines the direction of relative slip of the missing block.

Fig.(6.8)

10) Topographic characteristics such as drainges that are controlled by faults and fault scarps.


Northern Tibet: Triangular facets are fault scarps

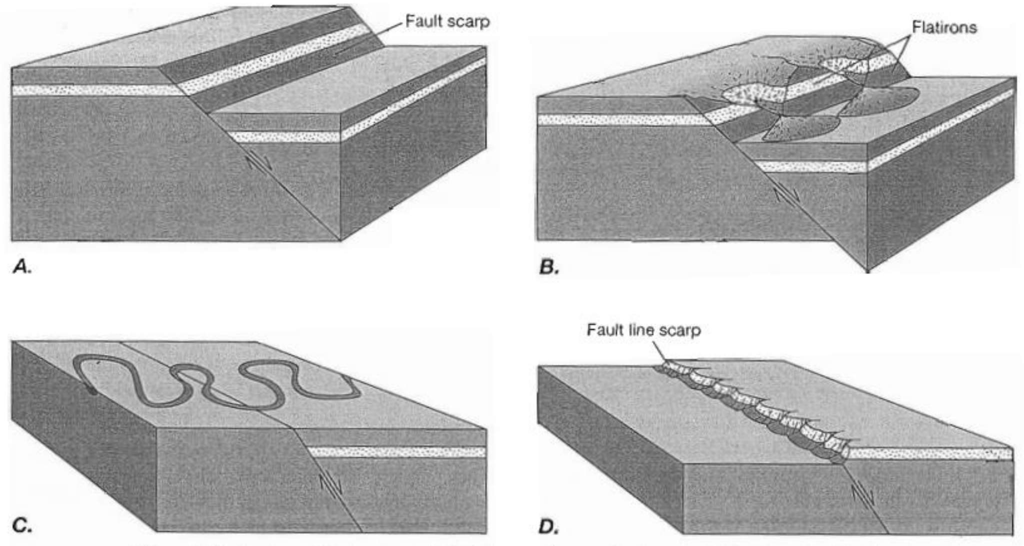


Figure 4.12 Erosion of fault scarps. *A.* Faulting produces a fault scarp. *B.* Erosion of valleys in the fault scarp produces flatirons. *C.* Erosion wears away the thin resistant layer in the topographically high footwall block and levels the topography. *D.* Erosion reaches the level of the resistant layer in the hanging wall block. More rapid erosion in the less resistant layers in the footwall block leaves a topographic step, a fault line scarp.