EXTENSION SYSTEMS

Extension systems are zones where plates split into two or more smaller blocks that move apart. To accommodate the separation, dominantly normal faults and even open fissures lead to stretching, rupture and lengthening of crustal rocks. At the same time, the lithosphere is thinned and the asthenosphere is upwelling below the necked lithosphere. Decompression during upwelling of the mantle results in partial melting. The produced basaltic magma is injected into the fissures or extruded as fissure eruptions along and on either side of the splitting linear region (graben and rifts). This mechanism, coeval lithospheric stretching and accretion of buoyant magma, is called rifting. It is called seafloor spreading once a rifted region becomes a plate boundary that creates new oceanic lithosphere as plates diverge from one another. The spreading centres shape elevated morphological forms, the mid-oceanic ridges, because magma and young, thin oceanic lithosphere are buoyant. Divergent plate boundaries are some of the most active volcanic zones on the Earth. Seafloor spreading is so important that it has created more than half of the Earth's surface during the past 200 Ma. Since the new continents drift away from the locus of extension, they escape further deformation and marine sedimentation seals relict structures of the early rift on either side of the new ocean. These two sides are passive continental margins.
The dominant stress field is extension. Bulk lithospheric rheologies control the development of large-scale extensional structures that can be classified as follows:

Continental lithosphere	
Narrow rift systems	East African rift system, Rhine Graben, North Sea
Wide extensional systems	Basin and Range
Passive margins	Bay of Biscay
Oceanic lithosphere	
Young oceanic basins	Gulf of Suez, Read Sea
Mid ocean ridges	Mid-Atlantic ridge
Back-arc basins	Philippines

In terms of plate tectonics, extension systems are associated with:

- constructive plate boundaries (oceanic and continental rift zones)
- destructive boundaries (back-arc spreading, marginal basins)
- intra-continental regions (graben, rifts and extensional basins).

Smaller-scale extensional settings include gravitational collapse of thickened crusts and continental margins, and local extension due to magmatic and salt domes.

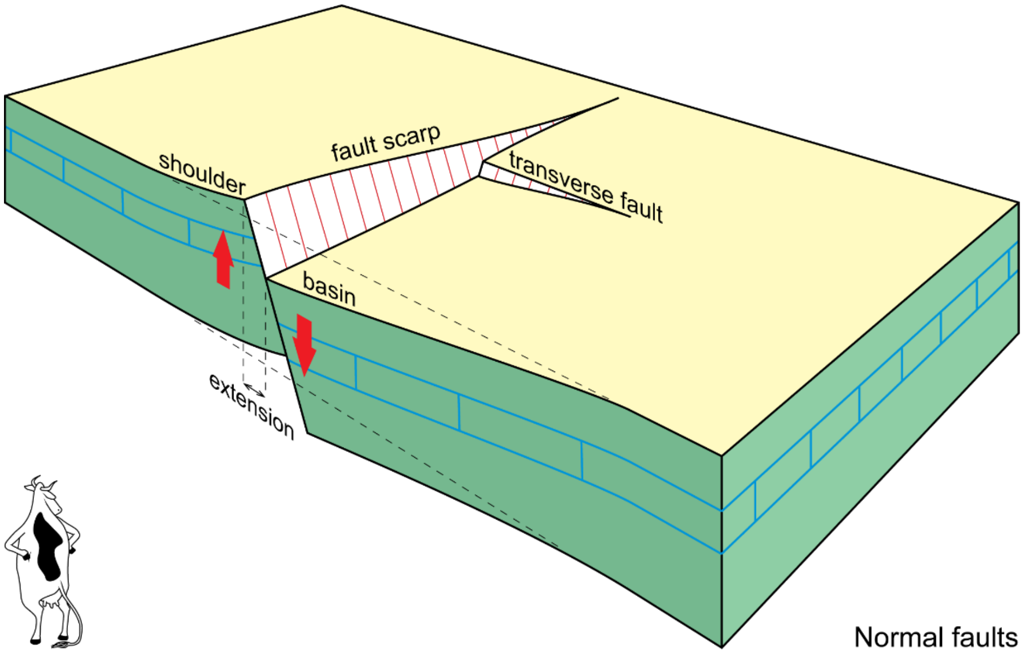
GEOMETRIC RULES OF NORMAL-FAULTING

Basic terminology

Many of the following terms are common to all fault systems.

Definition

A normal fault is a high angle, dip slip fault on which the hanging wall has moved down relative to the footwall.



Drilling through an extensional normal fault encounters a stratigraphic gap: normal faulting places younger rocks upon older rocks. Because of the separation of geological horizons that results from normal faulting, such faults are also termed extension faults.

Detachment

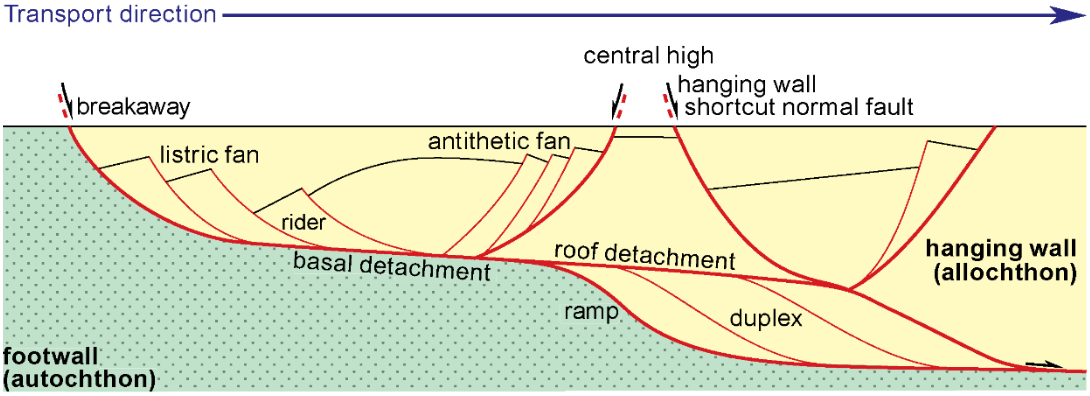
A normal fault dipping less than 45° is also called lag. In the modern literature, such shallow-dipping faults are termed detachment or denudation fault. A typical detachment has no root zone and follows a stratigraphic horizon. Confusion of detachment surfaces with thrust planes can be avoided from the superposition of younger rocks over older rocks.

Seismogenic layer

Earthquakes within extending continental lithospheres typically nucleate within the uppermost 15 km of the crust. Deeper in the crust, deformation is assumed to be ductile, taking place along shear zones. The horizontal boundary between the upper seismogenic layer and the underlying aseismic crust is considered to be a decoupling surface below which brittle faulting cannot grow.

Normal-fault system

Normal faults that formed and often interacted during the same extension event make together an extension (normal-fault) system. Master faults persist and have repeated movement over long periods of time. Associated faults with subordinate importance are secondary.



Fault and fault-block terminology of an extension system

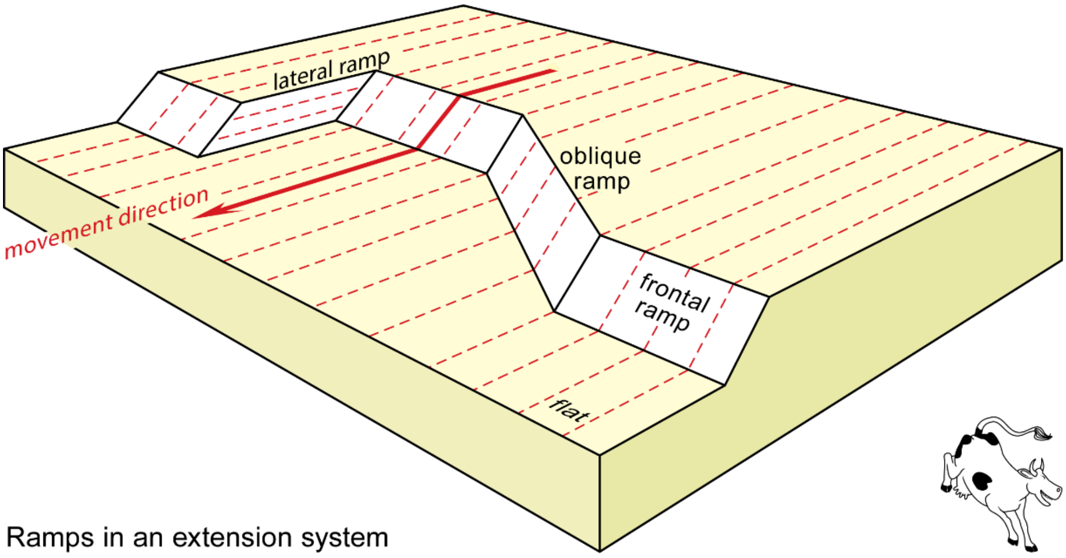
Transfer faults are lateral ramps linking or intersecting at a high angle neighbouring fault segments and involving strike-slip components to transfer displacement from one fault segment to the next.

Fault trajectory

Traces of normal faults on maps and in profile are characteristically irregular and discontinuous.
In cross section, normal faults show a great variety of shapes. Planar faults have a constant dip with depth. Some normal faults are curved. Listric faults are concave upward, i.e. flatten with depth. Antilistric faults steepen with depth. Where normal faults affect a set of nearly horizontal bedded rocks, they generally follow a complex staircase path made up of ramps that link alternating flats.

Flat

The flats are where the hanging wall slides along relatively weak, sub-horizontal boundaries or mechanical discontinuities also called detachment planes. Sedimentary rocks, such evaporites and overpressured shales, often behave as detachment levels. Friction along detachment plays an important role in the distribution of normal faults developing in the hanging wall, spacing being larger above soft flats than above frictional ones.



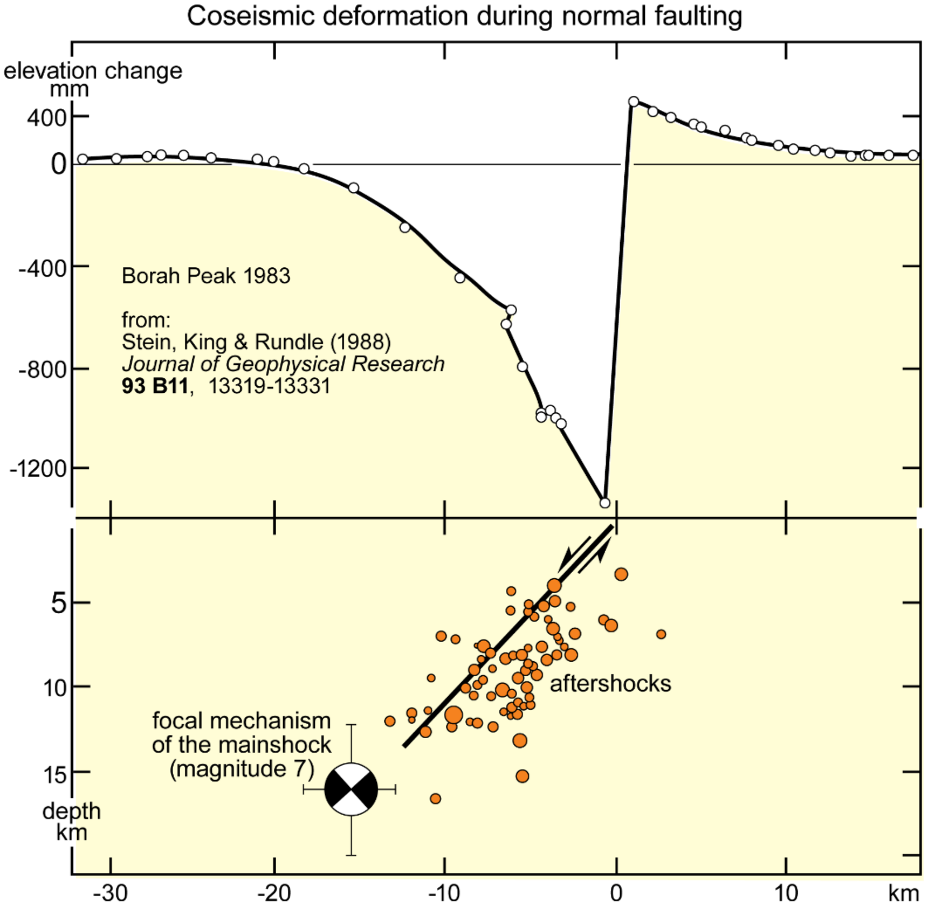
Ramps in an extension system

Ramp

The ramps are steeper segments where the fault plane climbs through the stratigraphic sequence typically at around 60° to the horizontal for normal faulting. Ramps do not necessarily strike perpendicular to the movement direction; they are also found oblique or parallel to the transport direction (lateral ramps).
Adjacent ramps merging with the same upper and lower flats form an extensional duplex.

Footwall, hanging-wall

Detachments commonly separate an undeformed footwall from a deformed and/or faulted hanging wall.



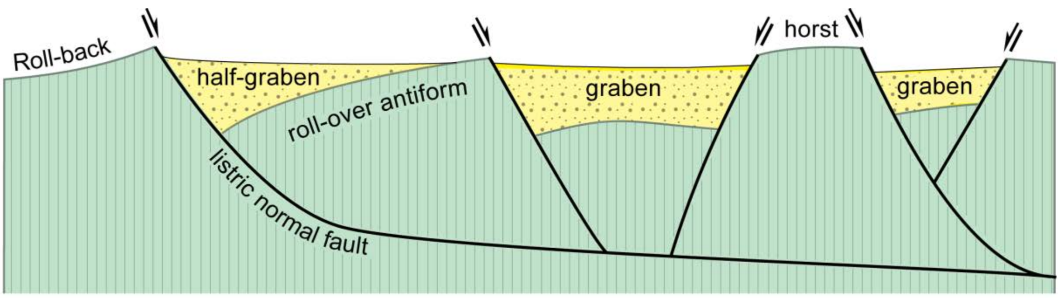
Geodetic measurements on the earth's surface before and after a major normal fault earthquake show that fault slip is accomplished by both hanging wall subsidence and footwall uplift.

Synthetic / antithetic faults

The hanging wall of a major normal fault often shows subsidiary normal faults either dipping in the same direction (synthetic faults) or in the opposite direction (antithetic faults).

Graben / Horst

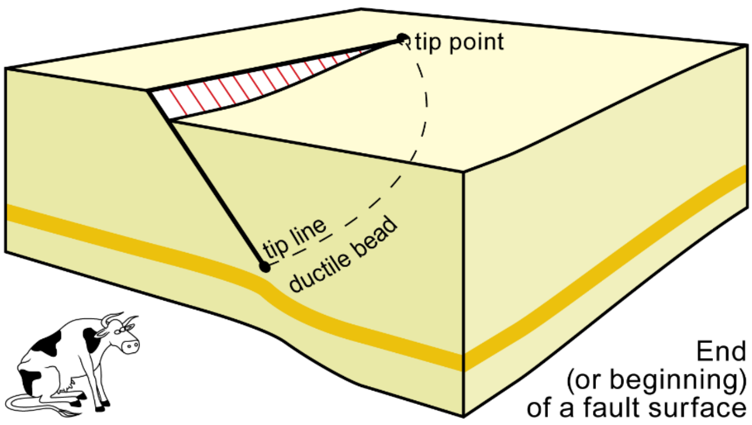
A down-dropped, hanging wall block bounded by conjugate normal faults dipping towards each other is a graben; a relatively elevated footwall block between normal faults dipping away from each other is a horst. Ideally, the growth rates of graben/horst-bounding faults are equal so that there is no fault block rotation and grabens and horsts remain symmetric. Lithospheric-scale grabens, called rifts, extend for long distances. A graben bounded by a single set of normal faults on one side of a tilted fault block has a triangular shape and is called a half-graben. Such basins host sedimentary wedges.



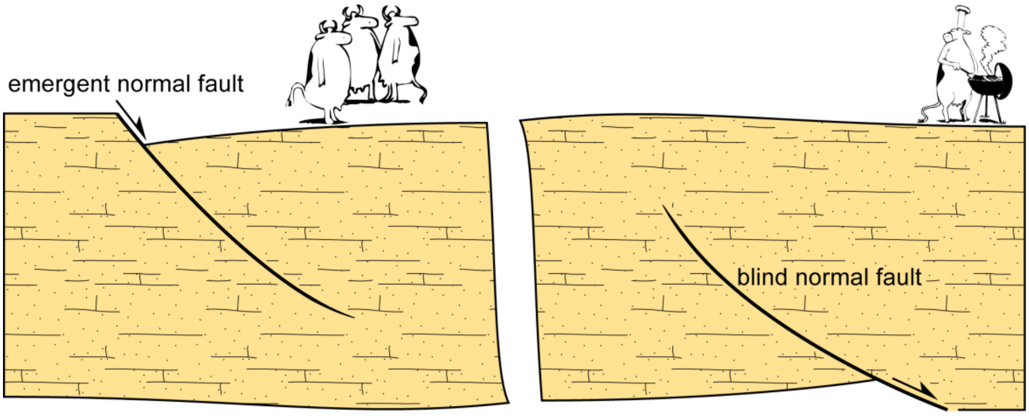
Main components of an extensional fault system

Tip line

The termination line of a fault is a tip line where the fault displacement has decreased to be accommodated by coherent deformation through the solid rock. In three-dimension, the termination line must be continuous and forms a closed line about the fault surface.

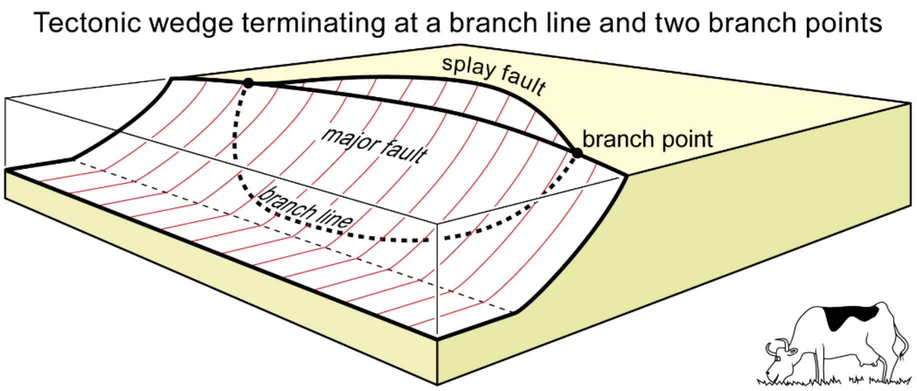


A fault plane cutting the earth surface is emergent. Conversely, a blind fault does not reach the earth surface.



Branch line

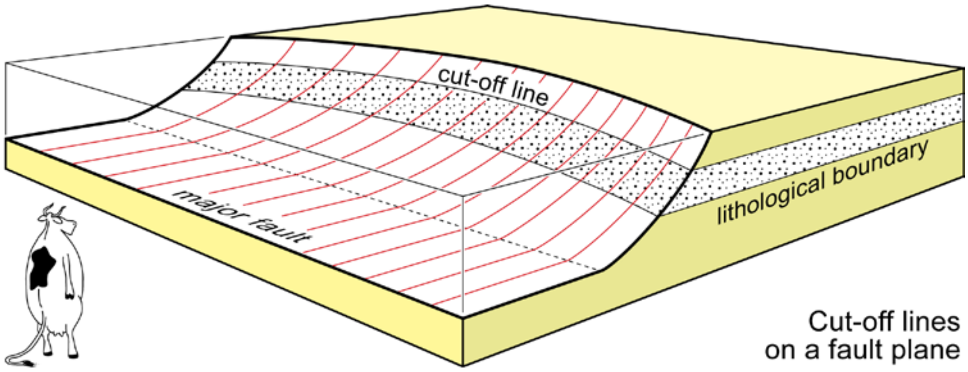
A branch line is a junction line where a fault splits into two fault surfaces of the same type.



Normal faults commonly die out in a set of smaller, subsidiary faults. Those are splay faults branching off from the main fault and forming an extensional fan that spreads the displacement over a large volume of rock. Splay faults are generally listric. Flexural flow or subsidiary faulting or volume change in transverse structures accompany fault movement where displacement is still larger relative to fault length (see tear faults, further down this lecture).

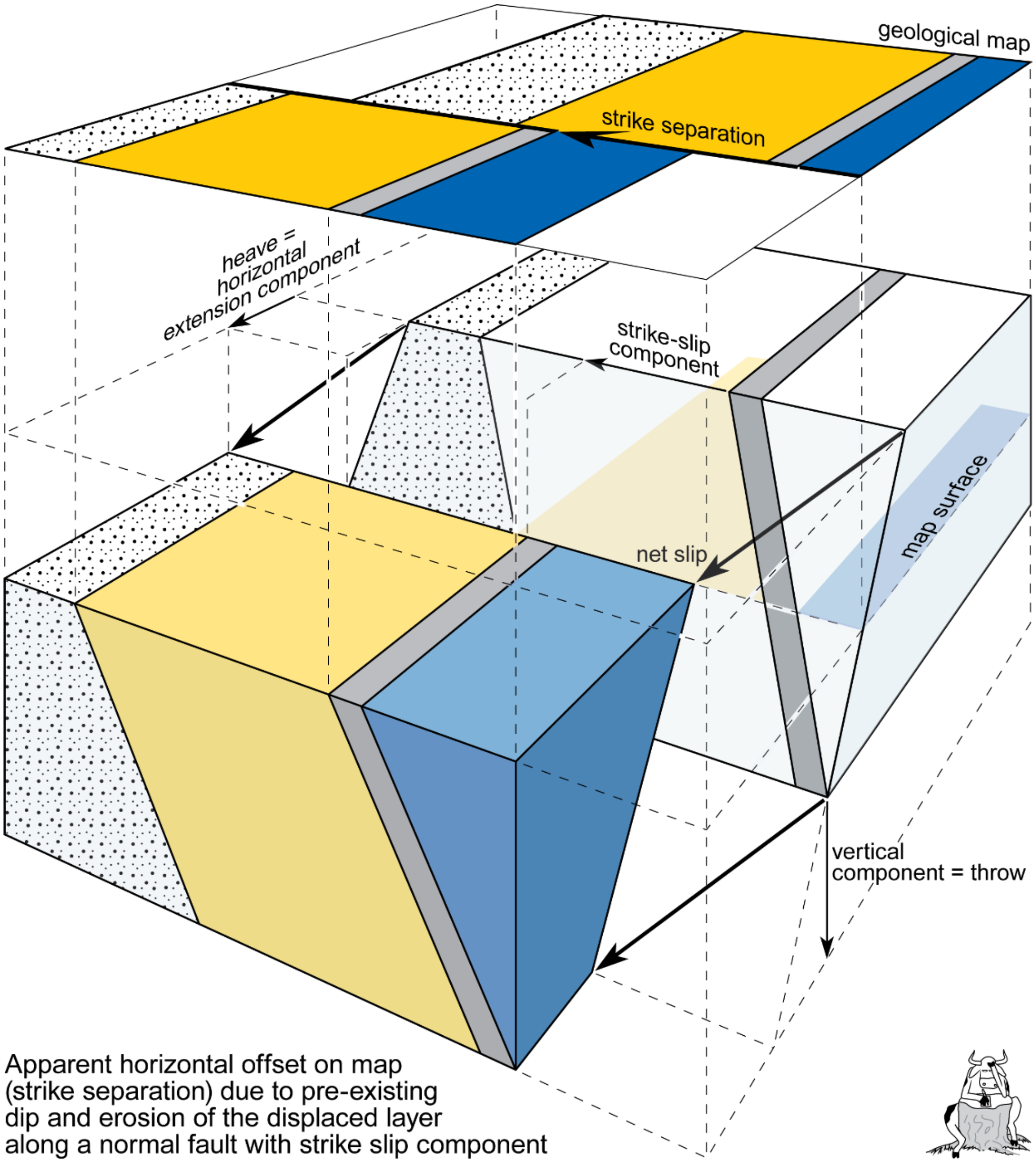
Cut-off line

The intersection between a particular contact (e.g. a stratigraphic surface) and a fault plane is the cut-off line. This line is a cut-off point on a cross-section.



Cut-off lines on a fault plane

The distance between the footwall and hanging-wall cut-offs of the same geological datum is the displacement. A slip vector with its magnitude (length) and orientation expresses this displacement. In a geological framework, the slip vector has a horizontal component (the strike-slip component) and a dip-slip component, normal or reverse according to the relative movement between the two fault blocks. The dip-slip component can further be decomposed into a horizontal component (the heave; elongation or shortening) and a vertical component (the throw; either uplift or subsidence).



General geometry of extensional faults

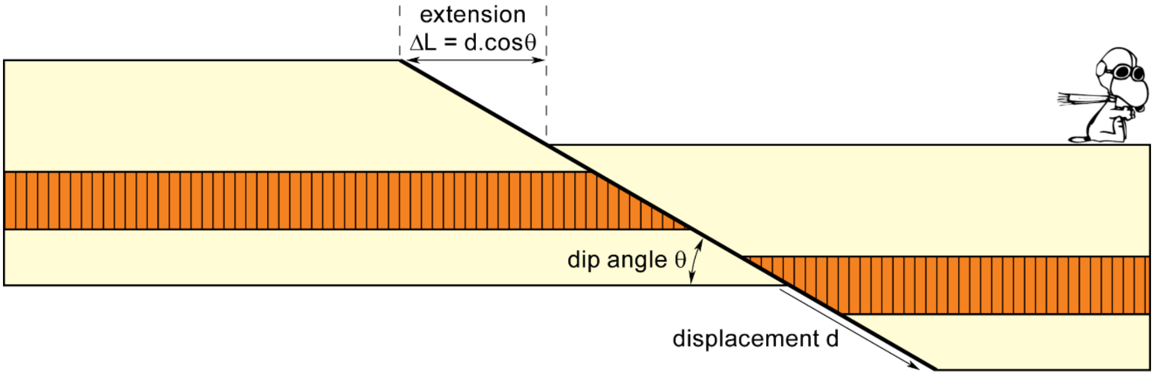
The orientation of the fault blocks may remain constant or may change as a result of faulting. The rotation during faulting depends on the geometry of the fault. There are three types of extensional faults.

a) Planar non-rotational faults
b) Planar rotational faults
c) Listric faults

Planar non-rotational normal fault

This is the classic normal fault type: a planar fault has constant dip; it involves vertical and lateral translations (throw and heave, respectively) but no rotation during faulting. Such faults control symmetrical grabens on dropped fault-bounded blocks between conjugate pairs (e.g. Rhine graben). A relatively high footwall block between adjacent faults that dip away from each other is a horst. Horsts and grabens commonly form because of the interaction between synthetic and antithetic faults.

Relationship between displacement and extension on a planar normal fault



In this simple case, fault strikes and dips are considered uniform. From a simple geometrical construction the change in length (ΔL) due to dip-slip displacement d on an individual fault whose dip is θ can be calculated. The relationship is:

$$\Delta L = d.\cos\theta\ .$$

The change in length of the region is the sum of the horizontal extensions on all the dip-slip faults. Assumptions limit the accuracy of extension amount calculated with this technique, although it can be used on limited segments of faulted crust with no deformation in the faulted blocks.
However this type of fault meets several problems:

1) How would such faults terminate at depth?
2) How do conjugate fault systems work? Are main faults contemporary or have they alternate movements?
3) They can only accommodate 30% extension.

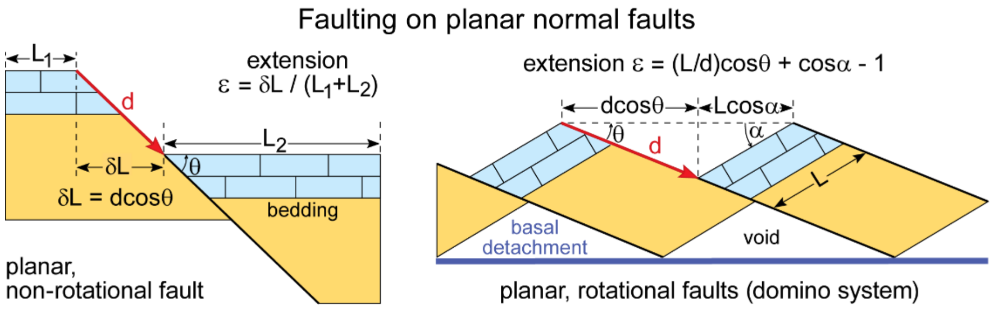
Planar rotational extensional faults

Parallel, planar fault planes and fault-blocks in between may rotate together about an axis roughly parallel to the strike of the faults in a way similar to the simultaneous tilting of a row of dominoes or a row of books on a bookshelf. Fault planes originally dip c.a. 60°, but this angle diminishes during the extension-related rotation. Rigid body rotation of the blocks significantly increases the horizontal component of slip on each fault, hence allows a larger amount of extension than non-rotational faults. The geometrical model of domino or bookshelf faulting assumes no penetrative deformation, pressure solution or bedding plane slip within the fault-blocks. Therefore, the angle between bedding and the fault planes remains constant; faults and fault blocks rotate simultaneously and at equal rate. Assuming that bedding was initially horizontal, the approximate amount of extension ε can be derived from the dips α and θ of beds and faults, respectively.

$$\varepsilon = \frac{\sin(\alpha + \theta)}{\sin\theta} - 1$$

Where ε = Extension in %
α = dip angle of bedding
θ = dip angle of the fault plane

Since natural examples are more complex, this equation gives only an approximation of the extension. However, the model predicts that shallow dipping normal faults may result from the rotation of initially steep fault planes.
Each fault block has its own half graben filled by sediments and associated volcanic rocks. Each fault has indefinite length or abuts against a transfer fault.



The model meets space problems at two levels:

(1) Each fault must have the same amount of displacement and tilting without any along-strike variation or there will be holes between adjacent blocks that do not rotate by the same amount.

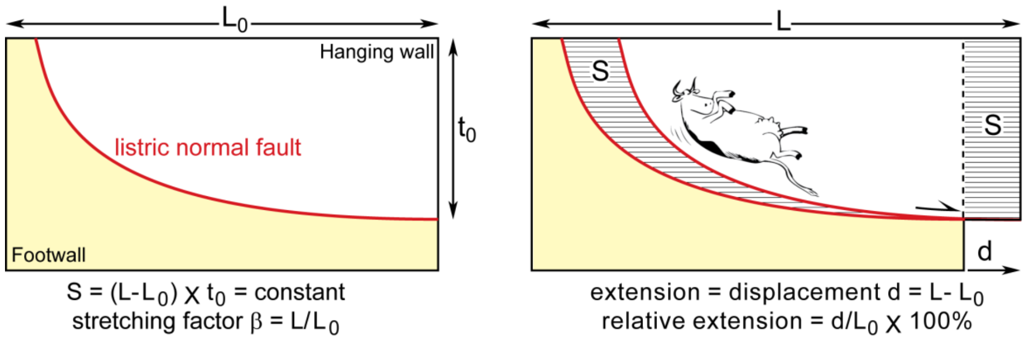
(2) At depth where the model predicts triangular gaps beneath tilted blocks and at the ends of the row. Small-scale faulting, brecciation, magma and more ductile deformation abrading corners and/or filling the gaps solve the problem.

Fault planes can rotate until at a very low dip and so can accommodate large extension. At low dips the fault blocks will lock and a new set of faults will initiate and begin to rotate. These new fault blocks will contain the old fault planes that will rotate passively. The new fault blocks will also develop their half graben.

Listric faults

Definition

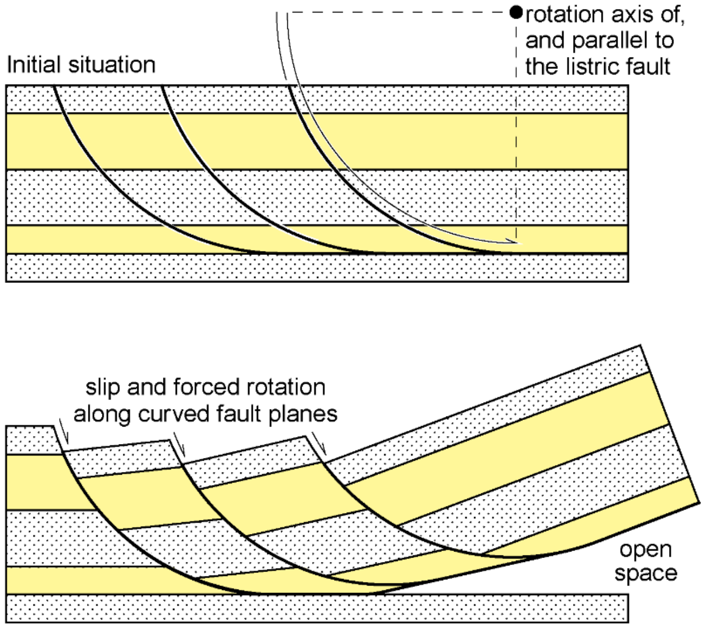
A listric (shovel shaped) fault is a curved, concave upward fault.



Such faults generally root in a gently dipping to flat detachment fault. The flat part of the fault usually develops along weak horizons of shale or salt above which strata, detached from their basement (i.e. thin-skinned tectonics), are displaced horizontally. Listric faults can accommodate unlimited extension.

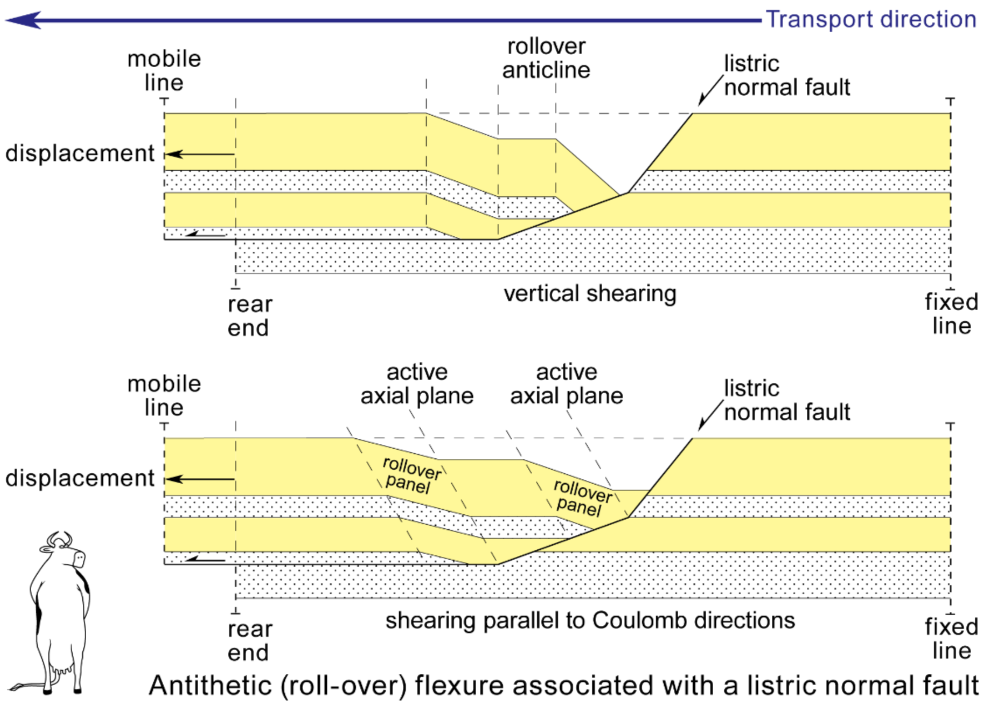
Rollover structure

Rigid fault blocks are geometrically forced to rotate if the bounding fault plane is curved. However, the hanging wall is forced to slide horizontally along the flat or shallow-dipping, bedding-plane detachment segment of listric faults. This translation opens a half-crescent-shaped gap between the hanging-wall and the footwall, above the curved, ramping part of the fault.



Model of cylindrical, listric normal-faults separating rigid blocks

Instantaneous collapse of the hanging-wall, due to gravity, closes the gap. With further extension, the hanging wall is deformed through faulting or rotation toward the master fault because rocks are not strong enough to support large voids and the hanging wall should maintain contact with the footwall. The result is a half-anticline, a rollover anticline in the hanging wall. As fault displacement increases, fault-bend folding (folding conforming the bent shape of the fault) continues and new growth beds (layers deposited while the fault is active) fill the resultant depression above the rollover fold.

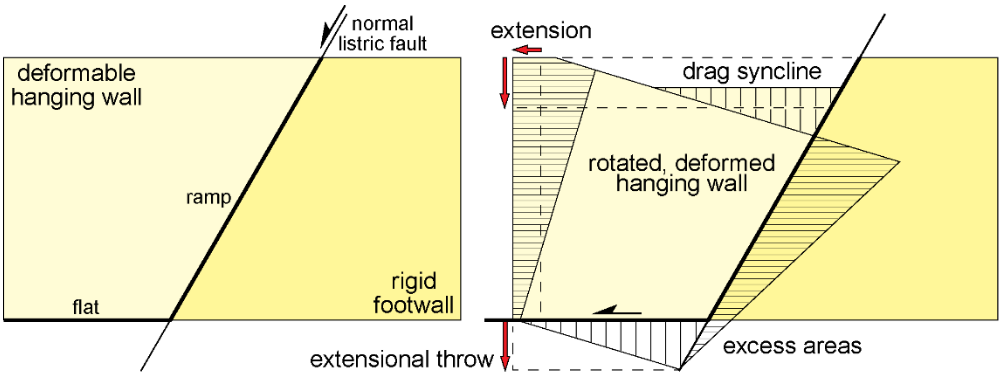


Antithetic (roll-over) flexure associated with a listric normal fault

The thickness of syn-extension sedimentary layers increases with dip of the rollover top toward the master fault. Consequently, the strata on the downthrown hanging wall are thicker against the master fault than the correlative strata of the footwall. The pre-growth and growth beds deform as fault displacement increases. Down warping of hanging wall strata toward a normal fault, also called

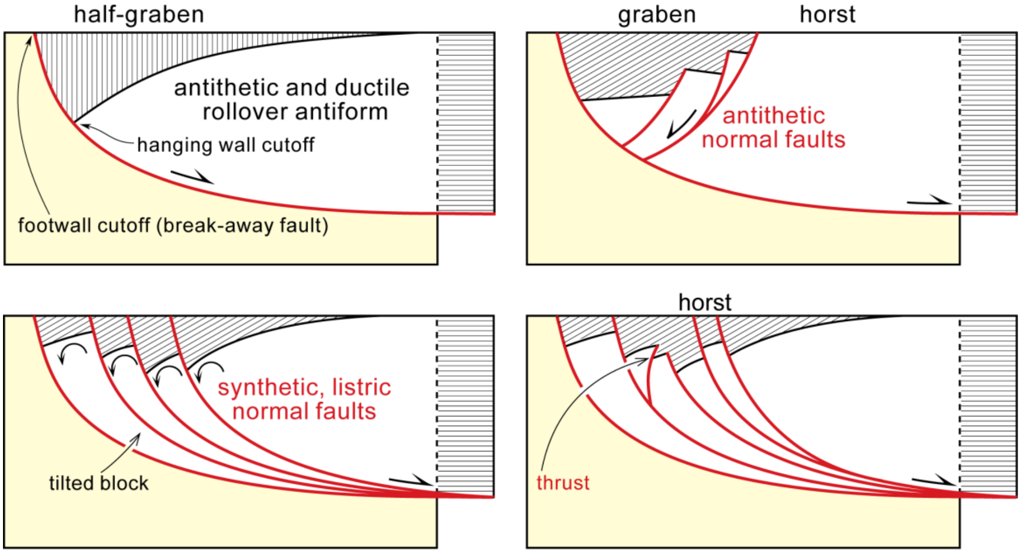
reverse drag, is widely taken as an indicator of listric fault geometry. Alternatively, a drag syncline may be the response of a deforming hanging wall against a strong footwall.
Note that the triangular shape of the half-graben over a rollover can be used to construct the change of dip of the associated listric fault; although this fault may appear nearly vertical on the surface outcrops, it is basically horizontal at depth. Older growth sediments dip at higher angles (can be up to vertical) than shallow growth sediments. Altogether, there is no simple geometric relationship between displacement and attitudes of bedding and faults if slip on a listric normal fault is involved.

Formation of a drag syncline
in the deformable hanging wall of a listric normal fault



Secondary faults

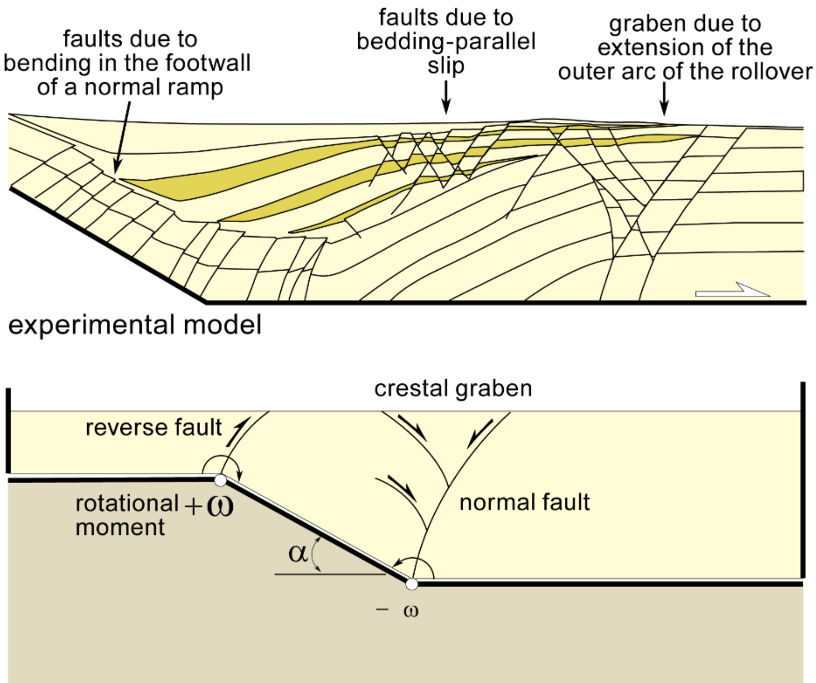
Synthetic and/or antithetic, planar and/or listric faults are necessary to accommodate the stretching of the hanging-wall rollover outer arc above the master normal fault in the absence of flexural slip or ductile rollover. Block displacement along these faults fills the potential void between the footwall and the hanging wall of the master fault. If listric themselves, secondary faults can reduce the rollover dip towards the master fault, favouring the development of secondary rollover structures with their corresponding ramp basins. The consequence is that listric faults occur in sets.



Models of collapse-deformation filling the gap between the hanging wall
and the footwall of a major listric normal fault

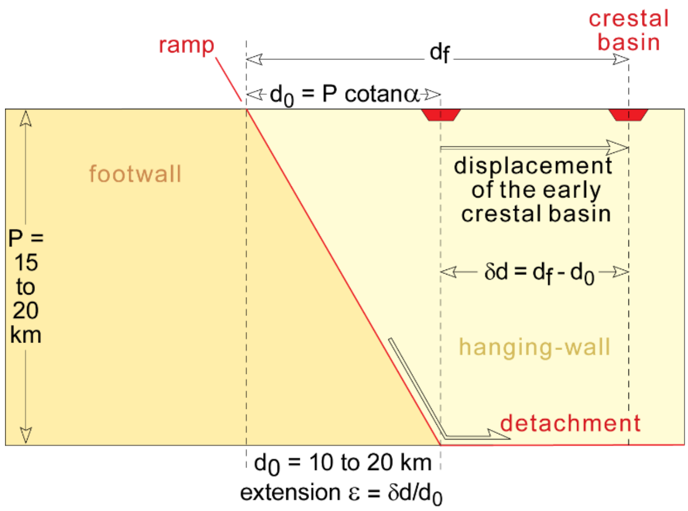
Analogue modelling

Analogue experiments show that early crestal grabens form in the hanging wall brittle layer almost directly above the connection line between the steep and flat segments of the fault. The original top surface of the hanging-wall block tilts towards the fault, which creates a half-graben basin that commonly overlaps the early crestal graben.



Interpretation of the initiation of an extensional system
from: Burg *et al.* (1994) *Géologie de la France* **3**, 33-51.

If one can identify early, relatively narrow grabens on one side of an asymmetric, rollover basin, the relative displacement δd of the graben can be measured, hence permitting a good approximation of crustal extension.



With further extension, some of the initial faults temporarily or permanently lock. Renewed extension on the underlying detachment then dissipates on other faults. Locked faults and their associated sub-basins then ride piggyback on the master listric faults. Several sub-basins may form part of a larger basin.

Detachment faults and core complexes

In cases of extreme extension, normal faulting strips off the shallower rock layers to expose rocks that originally were deep enough to undergo ductile deformation under metamorphic conditions. The crystalline basement rocks occur in window-like outcrops surrounded by a mylonitic, flat lying detachment that evolved into a cataclastic fault: the so-called metamorphic core complexes.

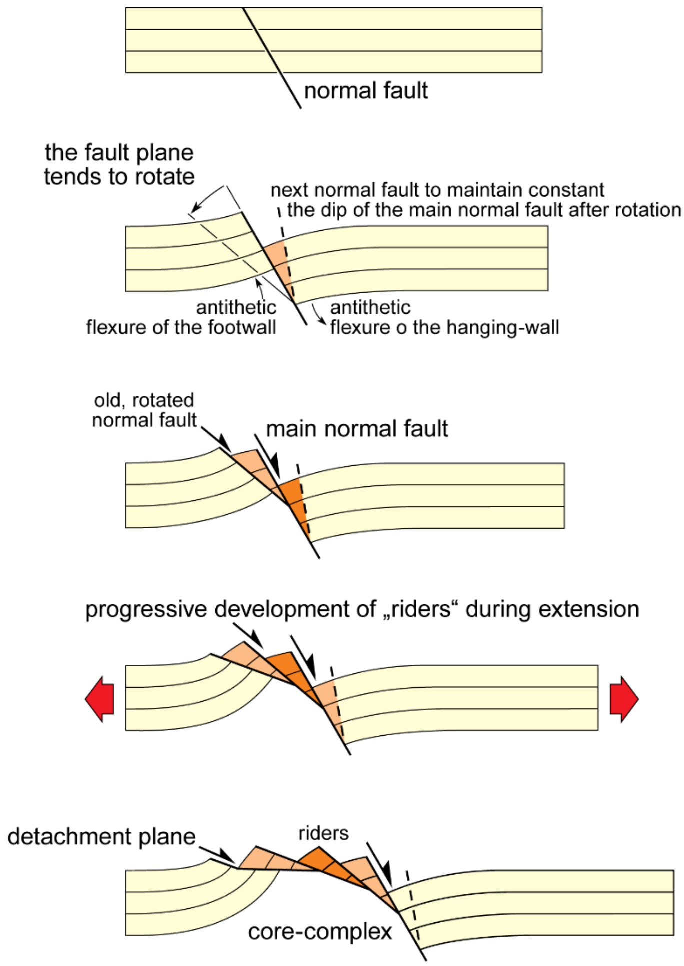
Detachment fault

Typical features of a detachment are as follows:

- It has no root.
- It usually takes place along a weak, stratigraphic horizon.
- Younger rocks will lie on older, often with a stratigraphic or metamorphic gap.
- Faults and brecciation are pervasive in the hanging wall and may be lacking in the footwall.
- Tight, overturned and recumbent, eventually faulted folds are common in incompetent strata.

Core complex

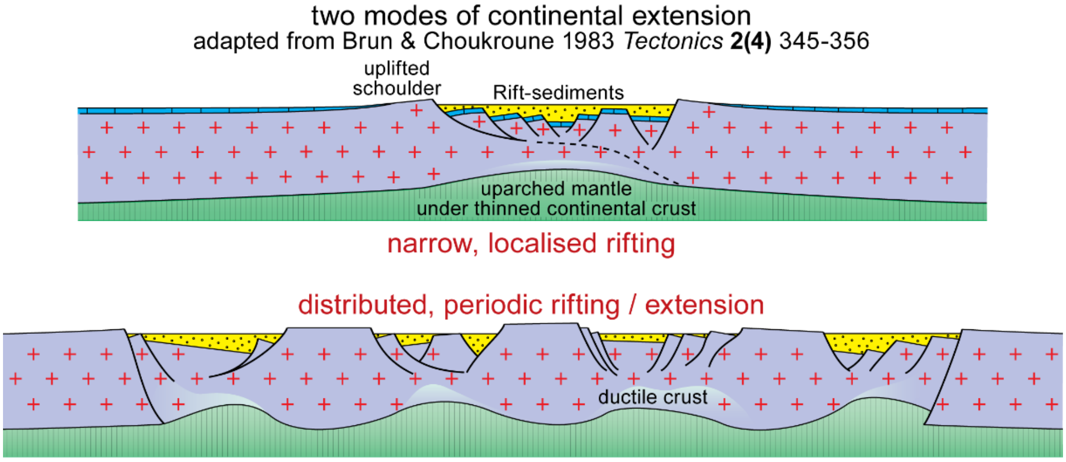
Plutonic and migmatitic rocks tend to rise when they become buoyantly mobile. Upward movement creates a gravitational potential for cover rocks to slide on a large-offset normal fault off to the side of the rising rocks. In response to tectonic denudation and unloading, the footwall core undergoes further isostatic rebound and uplift. The result of combined exhumation and extension is an elongated dome, the metamorphic core complex exposing high-grade rocks of the footwall of lower grade rocks, with a metamorphic gap along the strongly sheared, mylonitic bounding detachment.



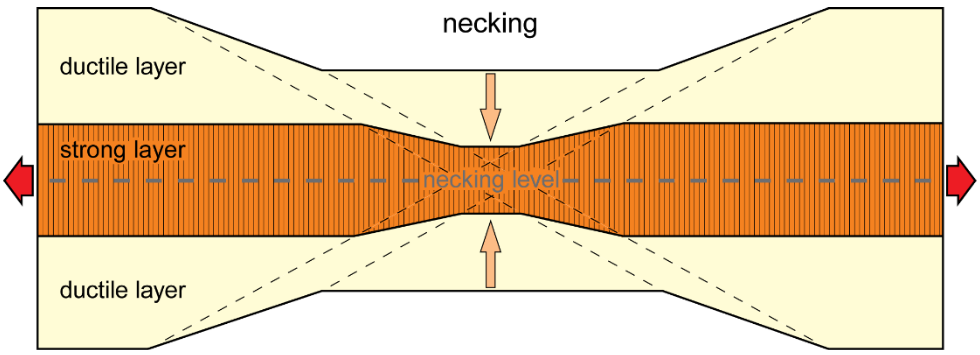
Development of a metamorphic core-complex
after Buck W.R. (1988) *Tectonics*, **7(5)**, 959-973

Crustal scale boudinage

During progressive experimental extension of two-layer, analogue systems, normal faults initiate with a steep attitude and delineate regularly spaced grabens and horsts in the brittle layer. Horsts remain virtually undeformed, yet occasionally rigidly tilted areas between the regularly spaced, extending graben sites. While normal faulting takes place in the brittle layer, ductile extension dominates the lower, viscous layer, which swells upward to compensate for the mass deficit arising in extended / thinned areas. The end product is like boudinage. Applied to geology, crustal scale boudinage is a mechanical instability expected in extended lithospheres, with ductile crust and mantle being elevated below grabens where lighter sediments replace crust. Heterogeneously distributed extension is consistent with the Basin and Range topography characterized by the alternating mountain ridge and valley landscapes in Nevada.



The necking level is the burial depth at which material remains stable. Rifting causes the material above the necking level to be displaced downwards while material situated below moves upward.



Sequence of faulting

During faulting at a ramp, the location of the ramp may change as the fault surface cuts in jumps into fault blocks. The result is often the stacking up of extensional, allochthonous sheets making up an imbricate zone or schuppen structure. Extensional duplexes may develop, characterised by a stack of horses that are progressively cut from the footwall block and added to the hanging wall block. The floor fault, which defines the bottom of the duplex, is the active fault, whereas the roof fault is never active at one time as a single fault. All parts of the roof faults undergo the same amounts of rotation, bending and faulting caused by extension on the master floor fault.

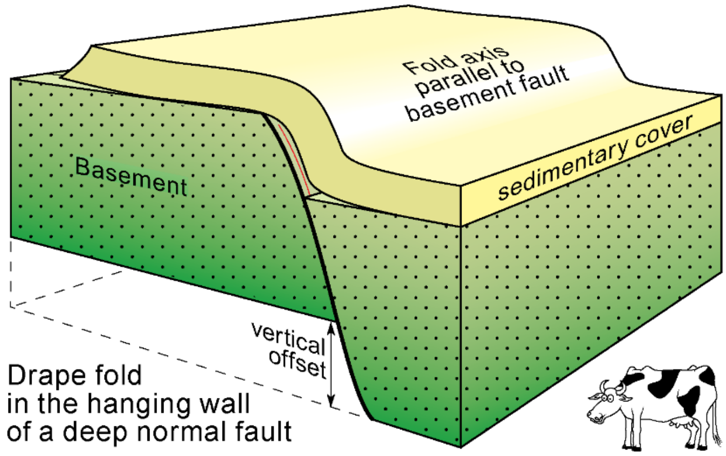
Relationship between folds and normal faults

Stratified cover rocks can be folded to cast basement fault offsets. The geometrical characteristics of drape folds of a passive hanging wall over the basement offset depend on the orientation of the fault plane with respect to the transport direction. It is worth emphasizing that such folds are not symptomatic for regional shortening.

Passive folds

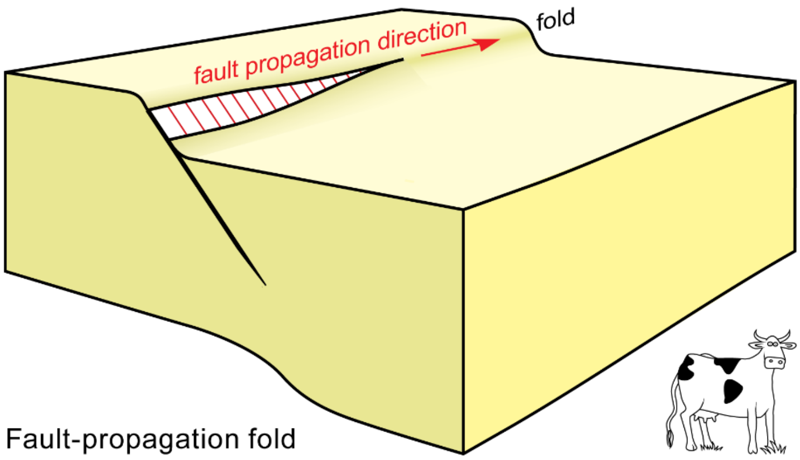
Fault-related drape and forced folds

Drape or forced folds may occur in sediments that cover and passively wrap basement vertical offsets due to blind normal faults. These fault-parallel folds may evolve into fault-propagation folds. Their amplitude depends on the vertical component of fault displacement.



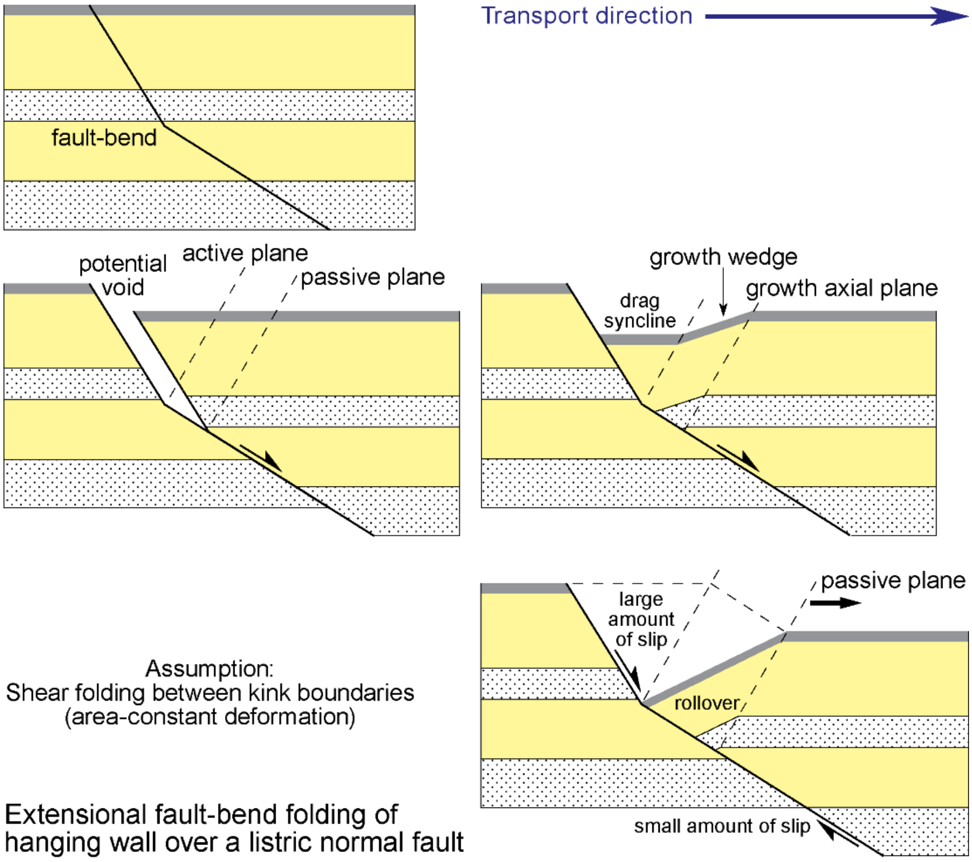
Fault-propagation fold – Normal drag

Fault-propagation folds are flexures in front of the fault tip while the fault plane grows; the expanding fault plane can later cut the flexure. The resulting geometry is similar to normal drag, which combines an anticline in the footwall and a syncline in the hanging wall.



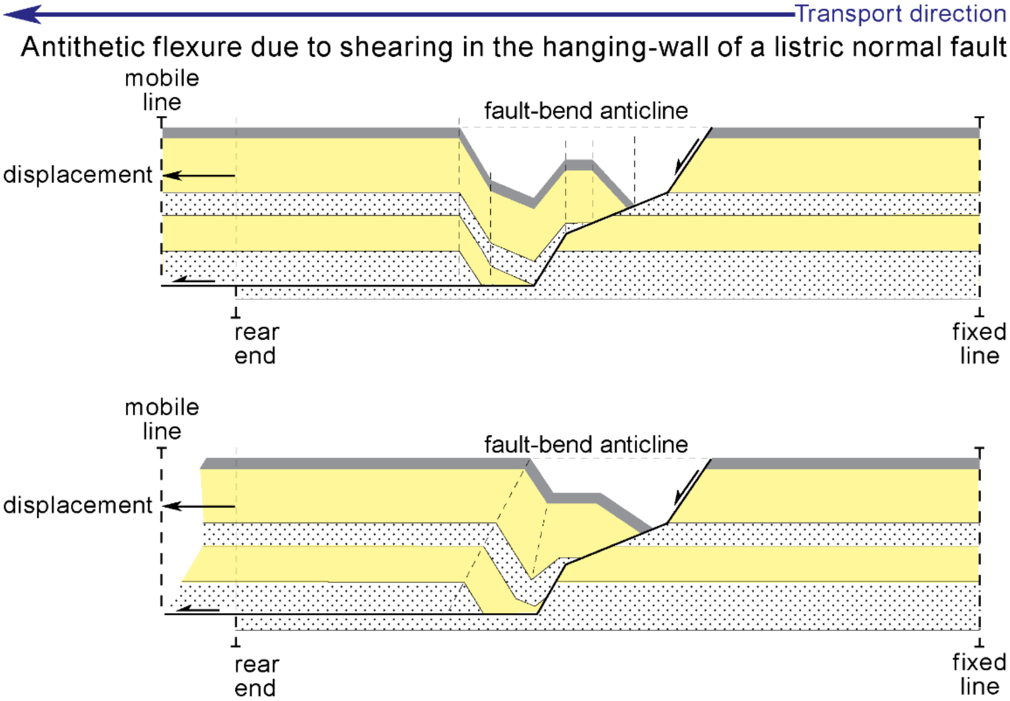
Fault-bend folds

Passage of the detached hanging wall over the fault bend of a curved normal fault induces a fold to fill up the potential gap between the hanging wall and the footwall. With increasing fault movement, hanging wall rocks fold continuously across the active axial plane, which is pinned to the foot wall at the fault bend, while the inactive axial plane, formed at the first movement increment, moves parallel to itself with the hanging wall.



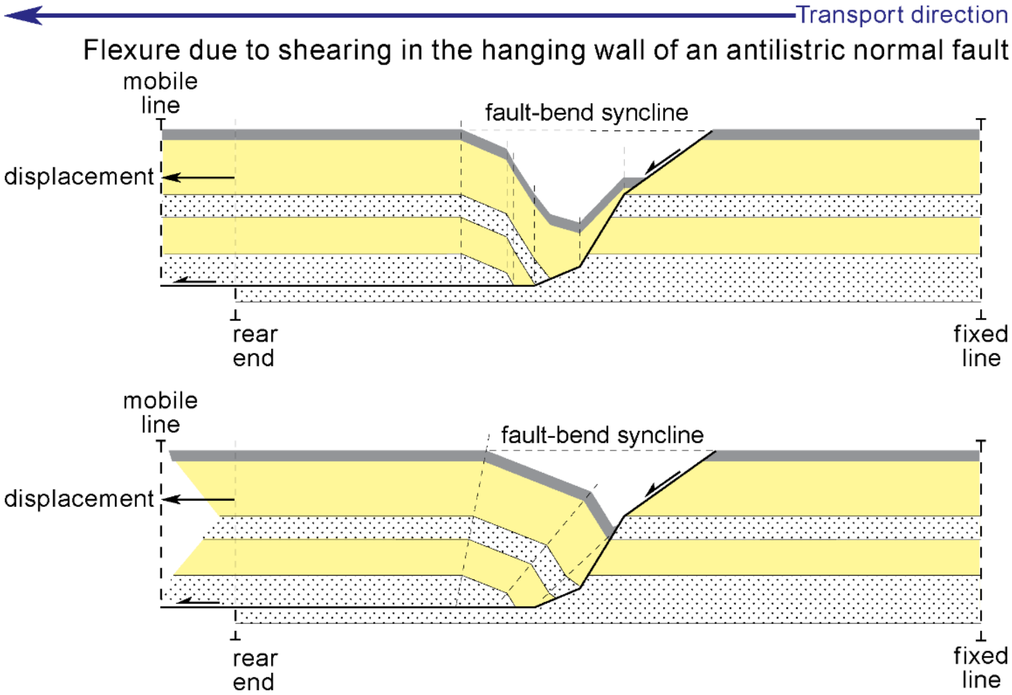
Extensional fault-bend folding of hanging wall over a listric normal fault

A rollover anticline is a gentle convex bending of hanging wall beds that developed to accommodate the upward concavity of a listric normal fault. Slip produces a fault-bend anticline, which is comparable to a rollover anticline.



Antithetic flexure with constant length in hanging-wall of a listric normal fault

Such fault-bend anticlines can be paired with fault-bend synclines due to antithetic rotation in the footwall. This fold pair is similar to reverse drag.
If a ramp is steeper than the bulk dip of the fault, slip produces a fault-ramp syncline.



Flexure with constant length in the hanging wall of an antilistric normal fault

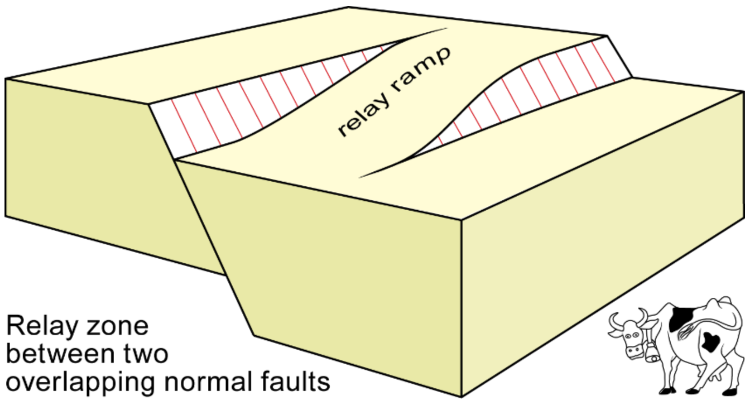
Transverse folds

Transverse folds are sub-orthogonal to major normal faults. They can be produced as local shortening structures in relay zones and as fault-bend folds above lateral and oblique extensional ramps.

Long-wavelength, low amplitude synclines with axes suborthogonal to the associated faults express the displacement gradient along the fault strike, from zero at the fault tip to a maximum somewhere along the fault length. Such folds may become cut by transverse normal faults that accommodate extension of the bent hanging wall. Antiforms maintain continuity at the fault tips and in zones of relay to neighbouring faults.

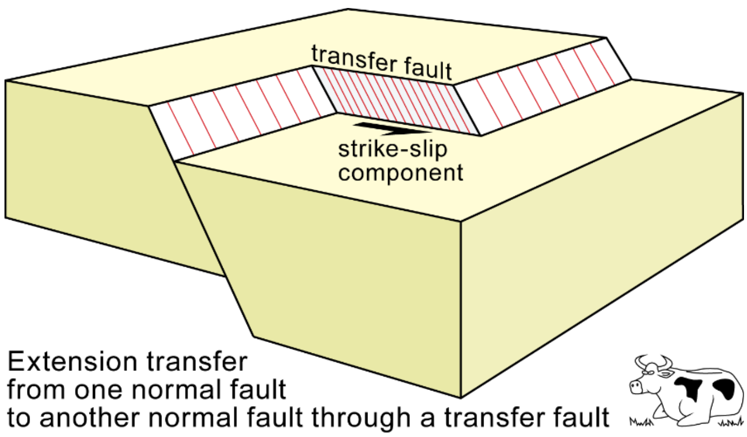
Accommodation zones: Relay ramps and transfer faults

In extensional systems rifting begins as a series of unconnected normal faults that die out at tip lines. Along-strike propagation during fault growth may link several of them. However, displacement on adjacent, overlapping and interacting normal faults takes up most of the regional extension. The interaction zone between these faults is an accommodation (transfer) zone within which extension is accommodated by folding or fracturing.



The general evolution in accommodation zones is:

1) Faults propagate along strike;
2) When faults begin to overlap, bedding is bent to form relay ramps in between. Relay ramps are areas of reoriented bedding between two normal faults that overlap in map view and often have the same dip direction. Bedding bending or tilting is the result of the decrease in displacement at the fault tips, so that this mode of deformation is sufficient to absorb a minor amount of differential displacements. Bent or folded accommodation zones constitute a common type of ‘soft linkage’.
3) With further faulting, the relay ramps become faulted. Transfer faults cut across the ramp, with a strike slip component, and connect the overlapping faults. Faulted accommodation zones constitute a common type of ‘hard linkage’.



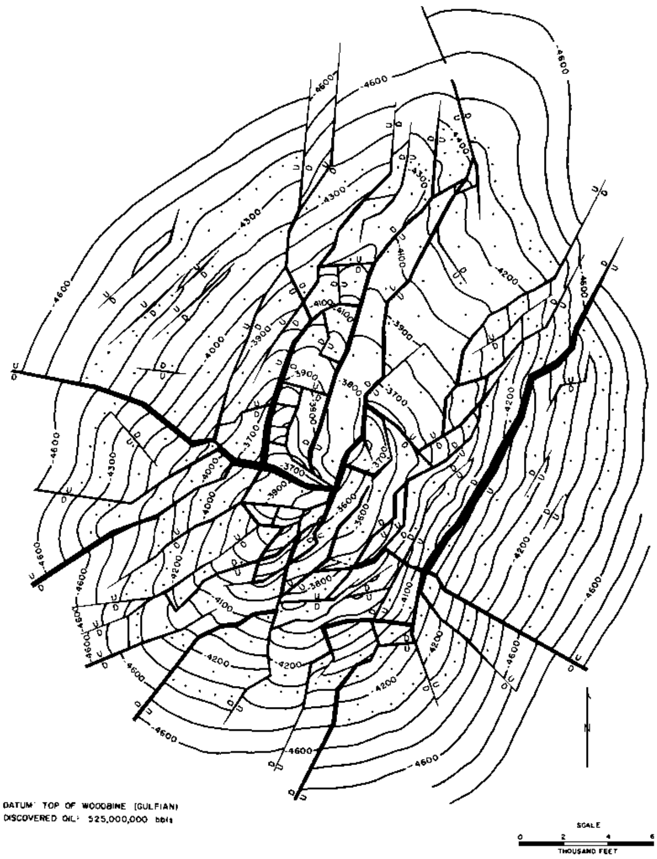
Experiments show that breaching often occurs through the propagation of the hanging-wall fault towards the footwall fault. Transfer faults play the same role as oceanic transform faults in transferring displacement from one fault to another, but differ in that all the movement planes are confined within the brittle crust. In addition, transfer faults may transfer displacement between adjacent regions undergoing different amounts of extension, different orientations of faults, or direction of tilting. Therefore, their geometry is quite variable. Predominantly strike slip systems are the only way to accommodate large differential displacements and strains, provided they are parallel to the extension direction. Transfer faults eventually may evolve into oceanic transform faults if rifting goes as far as to build an oceanic spreading centre.

Local normal faults associated with other structures

Localized extension zones are related to:

Domes

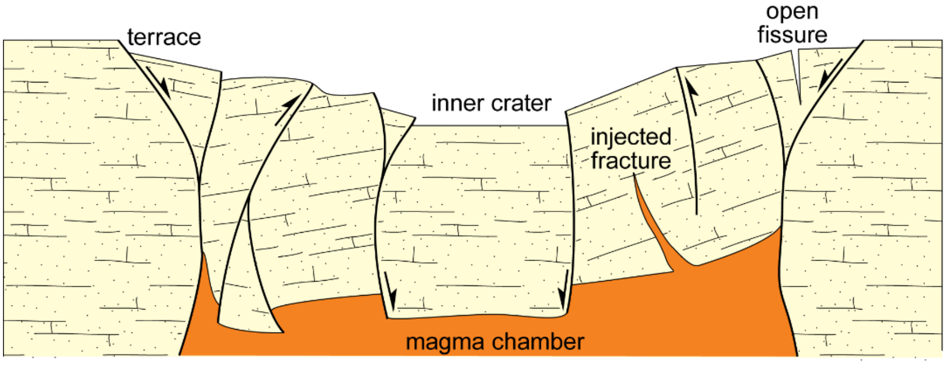
Structural domes may occur over salt and magma bodies. The local rise of deeper rock units leads to stress concentration and/or to strong stress gradients in the elevation zone. The subsequent radial and concentric normal faults generally stop at the edge of the dome.



Radial fault patterns over Hawkins salt dome, Wood County, Texas. Structure contours are on top of the Woodbine sand. (After E. A. Wendlandt *et al.* [1946], *Bull. Am. Assoc. Petrol. Geol.*, **30**.)

Ring faults in calderas

A caldera is a crater formed by a volcano collapsing into itself, usually because of low pressure in, or because of the violent removal of magma during eruption from, the magma chamber below (e.g. Santorini, Krakatau). Collapse of the magma chamber roof takes place along bounding, concentric faults (ring faults).



Faults around a caldera that develop successive outward

Two main types of faults exist in a caldera:
- (1) outward-dipping reverse faults mostly occur in the inner caldera and allow most of the collapse movement.
- (2) inward-dipping and curved normal faults form the peripheral rim of the Caldera in response to collapse along the inner faults.

Sand models suggest that deformation generally begins with broad sagging, followed by the formation of semi-circular arcuate or linear outward-dipping faults that propagate and interact around the caldera and typically form an overall polygonal structure. As subsidence continues, the caldera grows incrementally outward and progressively forms a series of concentric outward-dipping faults. Outer, inward-dipping normal faults form late as a result of subsidence along the inner faults. The depth and shape of the chamber affect the area of faulting, the symmetry of the caldera and the coherence of the subsiding block.

The faults at natural calderas determine locations and migration of eruptive vents, the degree of subsidence, the style of post-caldera resurgent magmatism, and the extent of hydrothermal circulation.

Folds

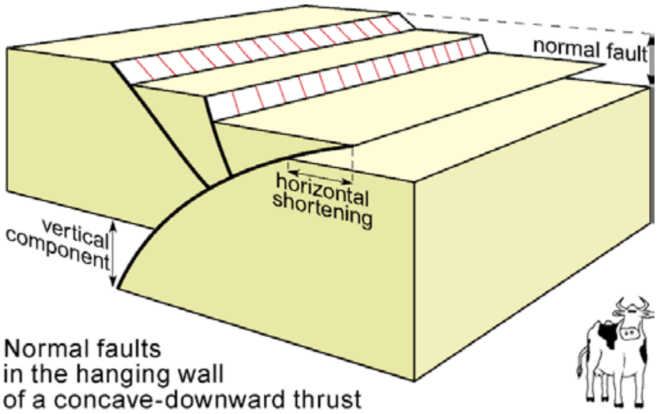
Normal faults accommodate extension in the outer arc of folds.

Pull apart

Pull-apart basins are rhomb-shaped depressions bounded on their sides by parallel and overlapping strike-slip faults and on their ends by diagonal (commonly 30-35°) transfer normal faults linking ends of the strike slip faults to the other strike slip fault.

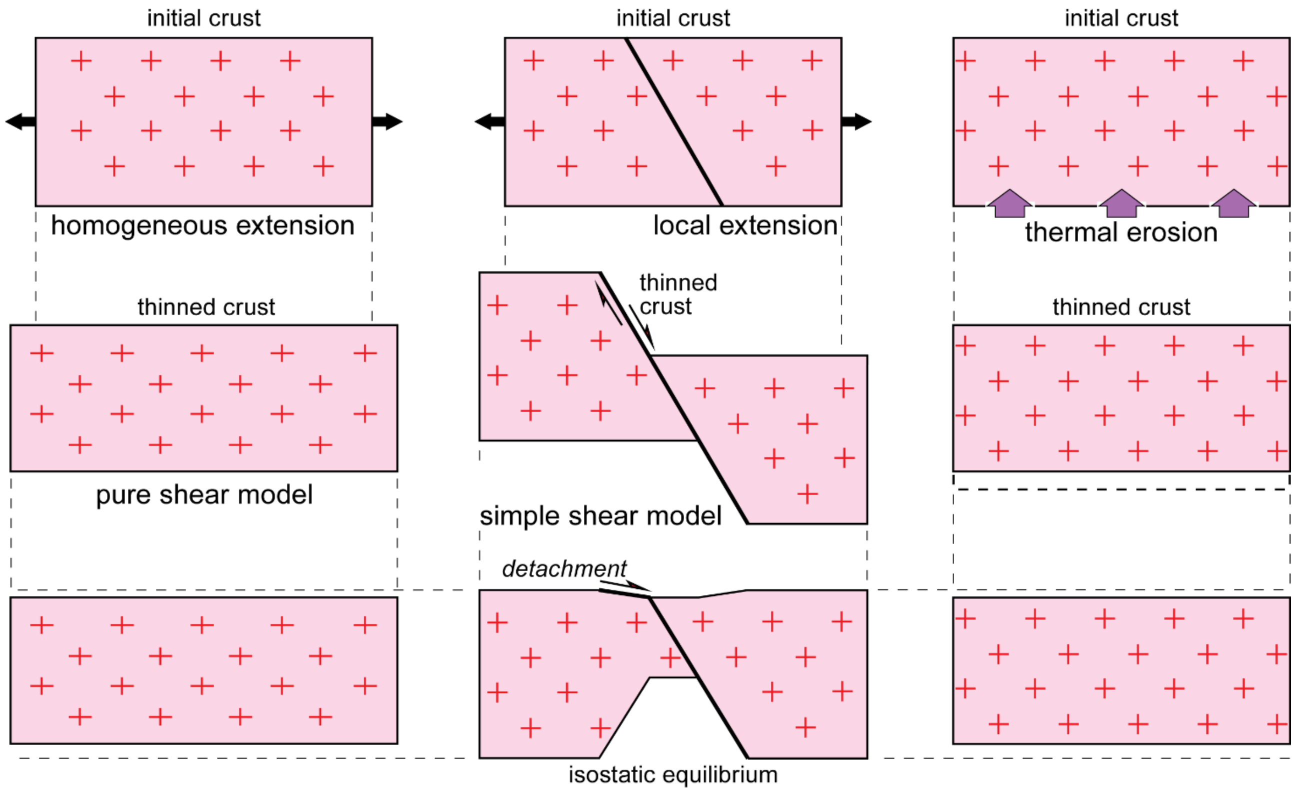
Thrust related normal faults

Normal faults and surface breaks are associated with local extension in the hanging wall of upward flattening thrusts. They allow a rollover type deformation that fills the potential gap that opens between the hanging wall and the footwall during thrusting. They are subparallel to the trend of the flattening ramp.



RHEOLOGICAL CONTROL OF EXTENSIONAL SYSTEMS

Models for lithospheric stretching range between two end members:
- Homogeneous pure shear, where crust and mantle extend homogeneously, and
- Simple shear where the lithosphere extends on a gentle to moderately dipping shear zone.



Three mechanisms of crustal (lithospheric) thinning

Structural characteristics of extended lithospheres (in particular passive margins) will obviously strongly depend on strength profiles. The base of upper-crustal normal faulting lies at or near the brittle-ductile transition in the crust. Beneath, the crust accommodates stretching by homogeneous ductile strain or movement on conjugate shear zones. Major flat detachments will lie at shallower depth along shallow brittle ductile transitions where thermal gradient are high. Furthermore, cold lithospheres and fast strain rates will tend to localise extensional deformation while warmer lithospheres and low strain rates will tend to broaden the extensional zone.

Pure shear model: instantaneous, homogeneous stretching

The pure shear (so-called McKenzie) model refers to a square marker in the pre-stretching crust becoming instantaneously a rectangle with the same area after uniform extension, which implies that crust and mantle thin equally and symmetrically. Lithospheric thinning allows passive upwelling of hot asthenosphere replacing the bottom thinned mantel directly beneath the upper surface basin. The lithosphere stretching-factor β is equivalent to the stretch in structural geology, i.e. it is defined as the ratio of the new length to the old length of a line. Thus:

$$\beta = L/L_0 = 1+\varepsilon$$

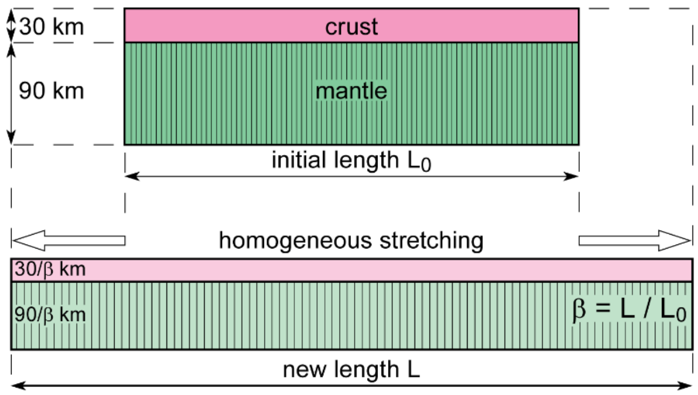
with ε the extension:

$$\varepsilon = \left(L - L_0\right)/L_0$$

Isostatic compensation accompanies stretching and produces an initial mechanical subsidence followed by a pronounced thermal subsidence phase. The amount of initial subsidence Si is the depth of the newly formed basin. It is given by:

$$S_i = d\left(1 - \frac{1}{\beta}\right)$$

where d is a complex factor incorporating the initial thickness of the crust and lithosphere, the densities of the mantle, crust and new basin sediments, the temperature at the base of the lithosphere and the coefficient of thermal expansion for mantle and crust.



Definition of the lithosphere stretching factor β

In reality, many of these parameters are not independent and the density of the new basin sediments mostly controls d. d ranges from c.a. 2.5 (air-filled basin) to c.a. 7.3 (sediment-filled basin).

The thermal anomaly due to instantaneous stretching decays exponentially with time, replacing relatively low-density asthenosphere with relatively high-density mantle lithosphere. This increase in bulk density causes a period of time-dependent thermal subsidence. In a simplified form the thermal subsidence at a time t is given by:

$$S_T\left(t\right) = E.\,r\left(1 - \exp\frac{t}{\tau}\right)$$

where E, the elevation to which the surface of the lithosphere sinks, depends on properties of the lithosphere, r on the stretching factor β and τ is the thermal time-constant of the lithosphere.

This two-stage model thus involves:

- An initial, syn-extension, isostatically controlled rapid subsidence.
- An exponential, post-extension subsidence caused by maintenance of isostatic equilibrium during cooling of the upwelled asthenosphere.

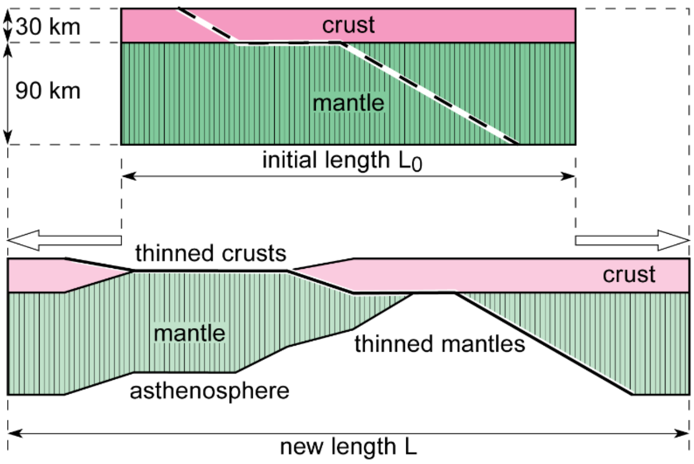
The model has provided an explanation for unfaulted basins known on the Earth. Calculations have shown that it gives a satisfactory approximation of basin history only for weakly stretched basins formed in a short time (β<1.5 over <30 Ma). Introducing time-dependent stretching, lateral heat flow and depth dependent stretching has refined the model. All of these pure shear models impart symmetry to the resulting basin. However, they do not predict the geometrical variability of extensional systems and provide no mechanisms for bringing mid-crustal rocks to shallow levels.

Simple shear models: asymmetric stretching

The simple shear (so-called Wernicke) model involves a gently-dipping detachment fault zone that cuts the entire lithosphere from surface to bottom. Owing to its oblique, shallow angle dip, the detachment zone offsets the thinned upper crust from the thinned lower crust, from the thinned lithospheric mantle and from the asthenospheric upwelling. Since the region of upper crustal extension is offset with respect to the region of deeper extension and asthenospheric upwelling, the zone of thermal subsidence is laterally shifted from the faulted upper crust. Isostatic compensation is again operating in this model.

Listric normal faults that sole out in the detachment cut the upper crust into tilted blocks that move against each other in domino-like rotations. The initial subsidence within the faulted upper-crust is:

- Greater than that in pure shear models.
- It is proportional to the extension and
- It is accompanied by a rise of, without extension at, the base of the lithosphere. This leads to moderate thermal subsidence as the lithosphere re-equilibrates to its pre-stretching thermal gradient.
- There is isostatic uplift on the side of the faulted crust, in the extensional transport direction, as hot asthenosphere rises into the thinned mantle beneath the unthinned upper crust. This is followed by thermal subsidence as the mantle cools. If erosion of the uplifted region occurs, then a shallow sag basin will be produced by this thermal event.



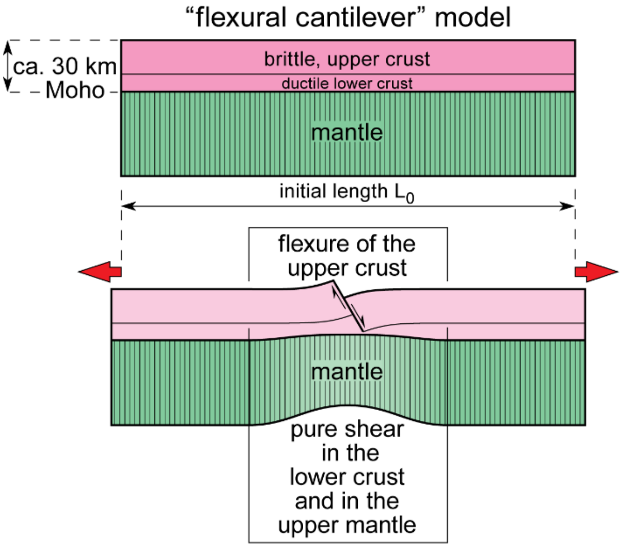
asymmetric stretching along a
lithospheric listric and shallow-dipping normal fault

The model predicts asymmetry at all scales.

- Upper plate margins are characterised by little stretched, therefore thick continental crust with narrow continental shelves and thin sedimentary cover because there is relatively little subsidence. They are structurally simple with weakly rotational normal faults. This region then undergoes only very minor thermal subsidence as the lithosphere re-equilibrates.
- Lower plate margins consist of substantially stretched, therefore thin continental crust with broad shelves and thick sedimentary cover because there is more subsidence. Since the lower crust and the mantle are dragged up along the detachment fault, the basement consists of exhumed middle- to lower crust, commonly overlain by remnants of the upper plate in tilt blocks.

Flexural cantilever models: depth-dependent extension

Flexural cantilever models combine listric fault-controlled simple shear in the upper crust with pure shear in the viscous lower crust and sub-crustal mantle, below a specified level of horizontal detachment. These models also incorporate the thermal effects, erosion of rift flanks and sedimentation in rift valleys. They predict syn-rift sediments to be thicker than post-rift sediments.



The term "flexure" refers to the bending of the elastic (on a geological time-scale) upper crust, assuming that this upper crust is an elastic beam or layer floating on a viscous, fluid lower crust and upper mantle. Displacement on an isolated normal fault causes flexural subsidence of the hanging-wall and flexural uplift of the footwall. The radius of the deformed region, i.e. the lateral extent of hanging-wall subsidence and footwall uplift, is controlled by the resistance of the elastic layer to bending (its flexural rigidity) and by the densities of the underlying and overlying materials. The flexural response to faulting affects a significantly larger region than the co-seismic elastic deformation (which is controlled by the size of the fault surface) and it is asymmetric because the overlying load is asymmetric sediment and/or water on the hanging-wall, air on the footwall.

Flexural models approximate the long-term fault behaviour to a summation of repeated co-seismic and post-seismic deformation around a major normal fault. Preferential loading of the hanging wall leads to increased hanging-wall subsidence and reduced footwall uplift.

Strain rate and thermal effects

Lithospheric extension increases the geothermal gradient by bringing the hot asthenosphere nearer to the surface, which induces significant alterations to the strength profile of the lithosphere.

If extension is slow, the geotherm may have time to re-equilibrate; hence, the base of the lithosphere moves downward and the burial depth of the mantle is reduced to compensate crustal thinning, which results in strengthening the lithosphere. It seems that slow extension should be self-limiting as rifting stops or migrates laterally or continues elsewhere when the mantle beneath the rift becomes sufficiently strong.

Conversely, rapid extension will lead to weakening since the temperature rise overwhelms the crustal thinning.

Numerical modelling predicts that fast extension rates are only possible for hot, thermally young lithosphere that will produce locally intense extensional deformation, with strain softening, leading to complete rifting of the continental crust and the formation of an ocean. For slow extension, with strain hardening, the locus of deformation is expected to spread laterally to involve a wider region of extensional deformation. Therefore, a link between rift width and strain rate should exist.

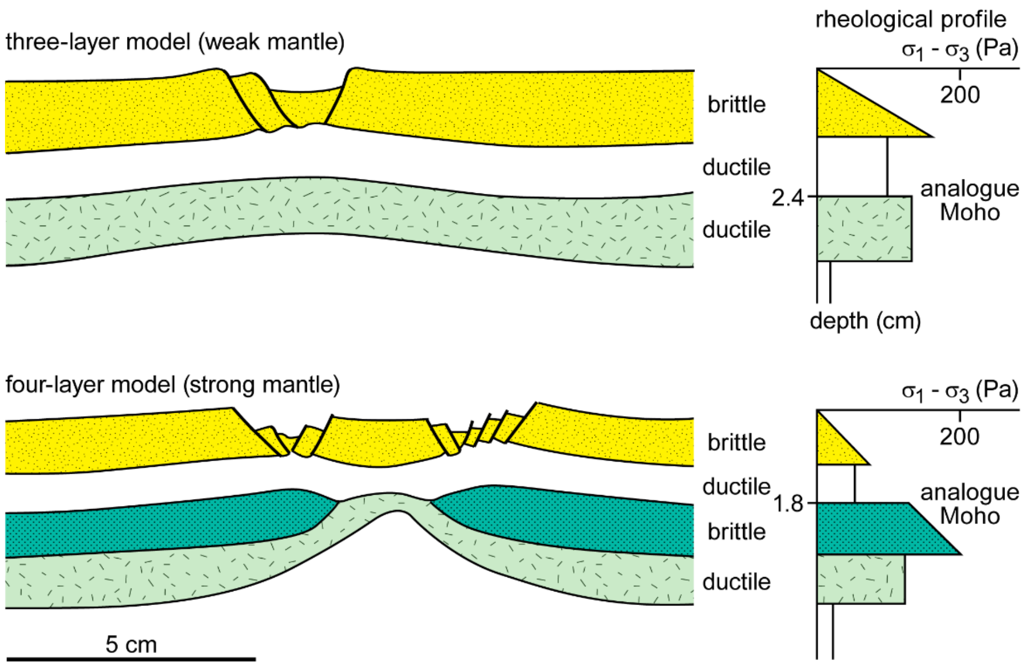
Localised versus distributed rifting

Extension may be localised within a single isolated rift (Rhine Graben, Baikal), contrasting with cases where extension is widely distributed in several, often regularly spaced and parallel grabens and horsts (Aegean Sea, Basin and Range). Analogue and numerical models show that the mechanical stratification of the crust provides the fundamental control on the width of the deformed region, fault spacing and, ultimately, the mode of extension, namely necking versus spreading. Since crustal temperatures strongly control the rheological profile and thus the elastic thickness and the crustal versus mantle depth of maximum strength, these two extension modes refer to the heat flow. Where the lithosphere is cold, rifts and adjacent deformation zones are narrow (ca. < 100km); where the lithosphere is hot, rifts and associated deformation zones are wide (>>100 km). Coupling between the viscous and brittle layers of the rheologically stratified lithosphere also plays an important role. This particular role may be expressed in terms of extension rates. In effect, a layer with a given viscosity can be a ductile decoupling level at low strain rates but becomes stronger with increasing strain rate. At low strain rate i.e. low coupling, the localised mode dominates. Deformation becomes more distributed with increasing rate, i.e. increasing coupling between ductile and brittle layers.

Necking

Metals and rock samples in tensile tests show that stretching, thinning and hence weakening of strong (competent) layers cause necking. Applied to the lithospheric scale, necking is called narrow-rift mode. Necking is a mechanical instability. It commences on any type of heterogeneity in the strongest layer, which initiates differential weakening and accelerating thinning in the weakest layers.
Lithospheric necking gives birth to narrow rifts, preferentially in a stable lithosphere where the heat flow is < 70 mWm-2 and the crust has a normal initial thickness (< 50km). Like necking in experiments on multilayers, rifting begins in the strongest parts of the lithosphere, usually the upper crust and the upper mantle. One or several of these competent layers become relatively thin in narrow regions of concentrated extension on intense normal faulting. Following this mode of extension, narrow rifts are characterised by large gradients in crustal thickness and topography. The bulk strength of a necking plate is reduced.

Extension modes of narrow rift zones in sand-silicone analogue models
after Brun (2002) *Geol. Soc. London. Spec. Pub.* **200**, 355-370



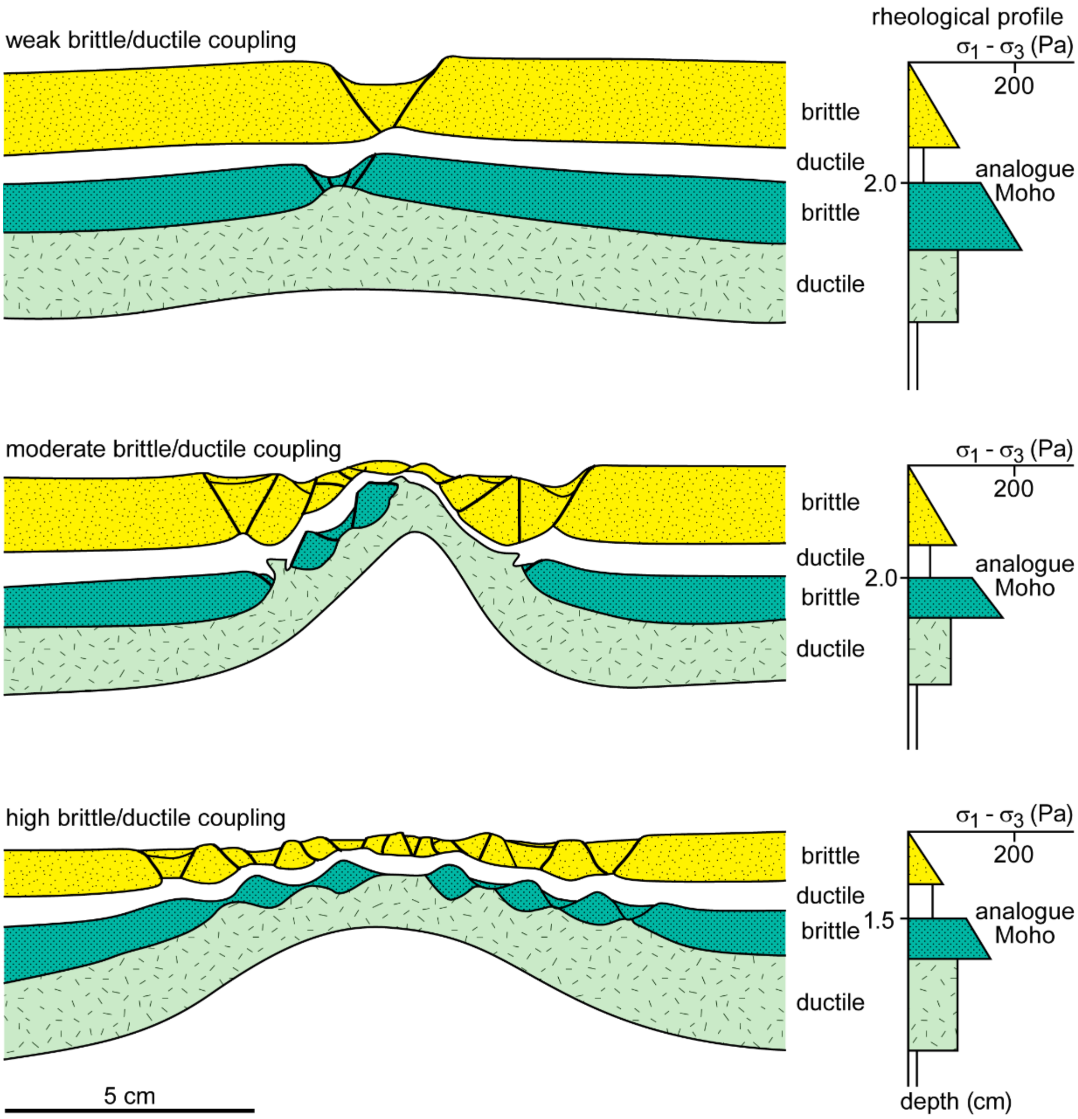
Distributed extension

Distributed extension (also termed wide rift-mode) often takes place in a lithosphere whose heat flow is > 90 mWm-2. Heterogeneously distributed but high extensional strain involves small lateral gradients in topography, a rather uniform thickness of the crust and lower-crustal and mantle thinning over a wide area. Two large-scale expressions depend on the rheological profile of the crust: distributed faulting and metamorphic core complexes.

Distributed faulting

Distributed faulting gives rise to a typical surface expression characterised by a large number of separated horsts and basins extending over a vast region (up to 1000 km). Such densely spaced normal faults with somewhat limited slip form when the strength ratio of the strong, brittle, upper crust to a weaker, ductile, lower crust is small, i.e., the lower crust is relatively strong. An example could be the North Sea.

Effect of the coupling between brittle and viscous layers
on the extension mode in sand-silicone analogue models
after Brun (2002) *Geol. Soc. London. Spec. Pub.* **200**, 355-370

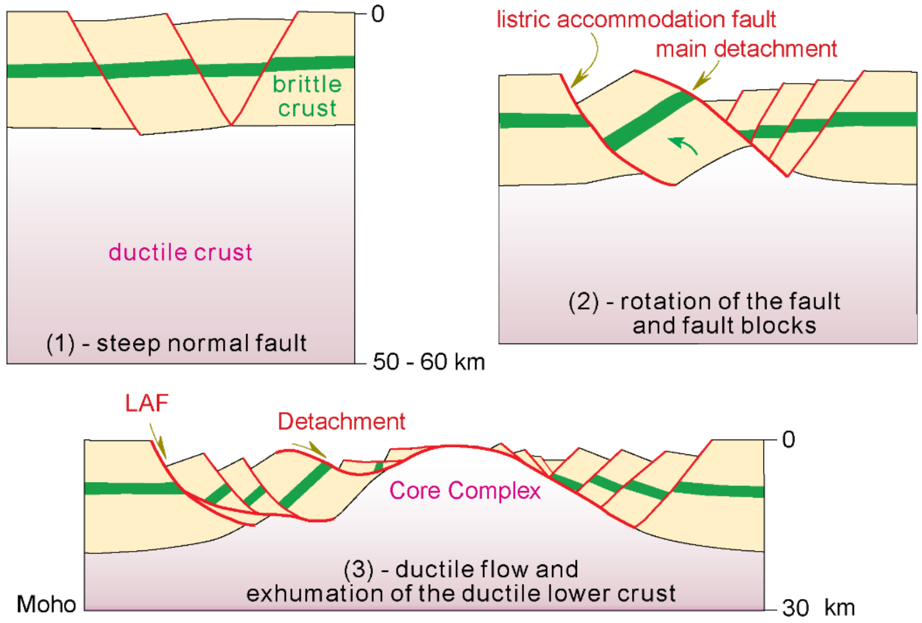


Metamorphic core complexes

This wide rifting-mode leads to the exposure of lower crustal rocks. Geological information shows that these rocks witness orogenic shortening/thickening to crustal thickness > 50 km during and/or

after cessation of convergence, before extension of the lithosphere. Early thickening buried much of the crust into high temperature conditions. The strength ratio between the brittle, upper crust and the ductile, lower crust is large, i.e. the lower crust is weak and flows easily. Stretching is then partitioned onto several detachment fault zones that accommodate large displacements and control the initial separation of the brittle crust on characteristic listric faults, eventually dissecting the upper crust and resulting in exhumation of the deep, medium to high-grade crust in metamorphic core complexes.
Two stages are usually recognized, rigid block faulting overprinting core complex formation. Examples are the western U.S.A. and the Aegean Sea. Metamorphic core complexes are typical for extensional destruction of the preceding mountain.

Development of a detachment system
model after Brun *et al.* (1994) *Geology* **22**, 319-322



Experimental extension of brittle/ductile systems produces similar domes of ductile layer rising beneath detachment systems. Localised deformation of the hanging wall is accommodated by a pervasive flow of the ductile crust that bends the detachment fault in a convex-upward, dome shape, resulting in the subhorizontal attitude of the inactive trailing part of the detachment zone. Whilst the ductile layer exhumes with further extension, one limb of the dome-like structure is bounded by the steeply dipping, active frontal part of the detachment fault zone. However, the other limb, due to block rotation, forms a footwall roll-under. As a result of this rotation, the brittle / ductile boundary is exhumed, giving a geometric marker to estimate the minimum relative displacement required to exhume the ductile crust.

LARGE-SCALE ANALYSIS OF EXTENSIONAL SYSTEMS

General features

Faults usually are elements of systems associating many faults.
From regional examples, the evolution from a continental rift to a new ocean is believed to begin with a dome in a continent where the lithosphere arches, expands, thins and fractures the crust, ultimately creating a fault-bounded rift valley. As the crust thins the mantle rises (mantle upwelling) and decompression melting produces basaltic eruptions. Partial melting of the granitic crust may produce rhyolitic magma. While the continental blocks move away from the hot, uparched spreading-ridge, the rifted margins begin to subside because the lithosphere cools and becomes denser. With continued extension new oceanic lithosphere is produced along the ridge between two passive margins. This

process is so important that more than half of the earth's surface has been created by volcanic activity along divergent boundaries during the past 200 million years. Implications are that continental rifting and the formation of new oceanic lithosphere involve a combination of magmatic and tectonic processes.

Extension systems

There are two major types of normal fault systems: Those resulting from local conditions, where gravity is the driving force, and those responding to far-field tectonic stresses. In gravity tectonics, listric faults are dominant and may affect the upper crust only. In tectonic extensional systems, normal-fault systems involve major planar faults that affect both sedimentary fill and basement of sedimentary basin.

Gravity-driven fault systems

Gravitational collapse of an elevated region, for instance a prograding delta, a high fault scarp or an unsupported sedimentary wedge is the typical example. In these cases listric faults develop without basement extension. Listric faults are common, link and sole out in a detachment across all scales. The detachment is at mid crustal level for larger faults and in incompetent layers (e.g. in shale or evaporite) for smaller faults. In crustal models, the rocks below the flat detachment are thought to be thinned by ductile mechanisms.
The structural characteristics of thrust-sheet emplacement through gravity gliding are the following:
(1) Basal thrust faults are listric and their trailing edge cuts up-section towards the surface.
(2) The potential displacement of each thrust sheet exceeds the section length of the thrust sheet.
(3) There is not necessarily any lateral stratigraphical continuity between adjacent thrust sheets.
(4) Transport paths may cross.
(5) The thrust-sheet with structurally the highest stratigraphy moves first and has the potential to move furthest. This is called diverticulation. Some geologists regard this as the main criterion by which to recognise gravity gliding.

Two other mechanisms may produce listric systems:

(1) Distribution and compensation of basement extension within sediments above a weak, easy-slip ductile layer (e.g. salt).
(2) Differential movements (with or without basement extension) around growing diapirs.

Rifts: Plate divergence in continental setting

The early stages of continental break-up are accomplished by normal faulting and produce rift systems. Dominant features are normal faults, fissures, dikes and volcanoes. Faults develop as planar dislocations, which may or may not be linked and commonly dip 55-75°. Large basin-bounding faults may cut through the crust. Mechanical unloading of the lithosphere and consequent isostatic rebound produce flexural uplift of footwall blocks on both sides of the rift, the high relief shoulders. The major listric fault bounding a rift is the breakaway fault.

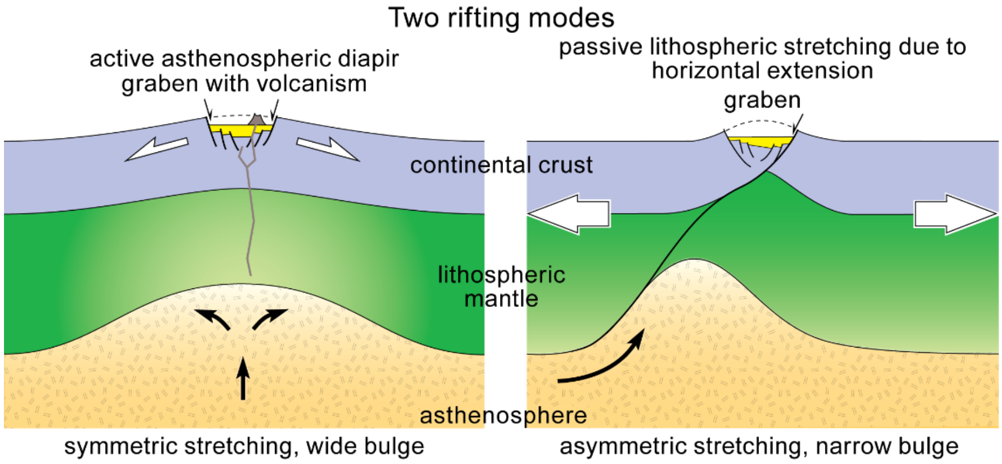
Morphology; Rifting and sedimentation

Active rifts are long and narrow depressions on the earth crust; they are characterised by seismic activity down to depths of ca. 15 km, high heat flow, and topographically elevated shoulders due to flexural isostatic compensation of the lithosphere. Some of the best examples comprise the Red-Sea, the African Great Rift Valley and the Gulf of Aden that meet in a triple junction in the Afar region. Rifting produces long and linear crustal depressions that are sites of extensive sediment accumulation. Major unconformities divide the general stratigraphy into pre-, syn-, and post-rift sequences. Rift characteristics are:
- A relatively narrow width (30-60 km), which is about the thickness of the rifted continental crust, irrespective of the length of the rift valley. Since active rift zones are initially elevated above the sea level, syn-rift sediments are often continental or lagoonal and strongly influenced by syn-rift

volcanism in places. Huge down-dropped blocks are sites of continental basins, which are filled by thick (> 1 km) clastic debris derived from adjacent high-standing blocks and deposited in alluvial fans. These sediments impart the relatively flat topography of the rift valley although they are deposited within an irregular pattern of partial deposition centers. Major lakes and rivers are common in this central depression.
- Steep and subparallel flanks are 3–5 km higher than the rift floor; this outstanding landscape feature results from cumulated normal-fault scarps, with the faults stepping down towards the central lowland. Some of these faults are very long but most of them relay each other, occasionally in an en échelon manner. Shoulder uplift arises from the mechanical unloading of the footwall block. Isostasy calculations indicate that the total relief of a rift wall is about 2.25 times the uplift. The amplitude of the uplift as well as the width of the rift flank uplift is determined by the elastic thickness of the lithosphere Te. The width of the shoulders varies from 80 km for Te=15 km (Rhine Graben) to 200 km for Te =50km (Baikal).
- Rifts are not equally deep all along their length. Thinning of the continental crust usually occurs on a series of sub-planar to listric faults with a throw of several km. Rifts bounded by such serialized faults are cut into segments and comprise an assemblage of half-grabens of varying polarity, arranged end-to-end or side-by-side, and linked through cross-rift accommodation zones. The dimension of the half-graben is determined primarily by the thickness of the seismogenic brittle crust, and is typically 25-100 km long and 20-50 km wide. Fault-bounded, long and narrow basins develop on the rift-parallel extensional blocks that rotate about a horizontal axis as the rift zone widens. The drainage patterns determine sediment input into the half-grabens. Their sedimentary fill is asymmetric, with coarse terrigenous sediments in aprons and fans rimming the fault escarpment and finer-grained sediments depositing on the moderately dipping counter-slope. Polarity reversals in half-grabens, due to the development of new faults as old systems become inactive, may complicate sediment distributions. These descriptions imply that many rifts are asymmetric, with one margin being higher than the other and the sub-sediment floor being tilted.
- Episodic pulses of extension create space for sediment accumulation at very fast rates. Stages of rapid mechanical subsidence are typically followed by periods of relative tectonic quiescence, when sediment supply fills the available space (called accommodation). Syn-sedimentary tectonics (block-tilting with sedimentary fanning, angular unconformities, stratigraphic gaps, etc) are very active. Alluvial and fluvial early-rift sediments pass upward into shallow and deep lacustrine facies (conglomerates, sandstones, coal, fine-grained mud) where intermittent lakes occur along the axis of the basin. As the rift widens, its floor further subsides and eventually sinks at sea level. At this stage, the sea that covers the basin is very shallow and may dry up. Evaporites precipitate in this environment. With continued extension the basin broadens and deepens, thereby becoming a narrow ocean in which marine strata bury the evaporites.
- Normal faults accommodating continental crustal pull-apart and accompanied by parallel dyke swarms and outpouring flows of both tholeiitic and alkaline basalts. Mounts Kenya and Kilimandjaro are big volcanoes that exemplify this magmatism. Partial melting of the granitic crust may produce rhyolitic magma. The bimodal association of acid and basic volcanic rocks is characteristic of within-continent rift systems. In fact, magmatic rocks show a great range and also include extreme compositions that are rarely found outside rifts such as extremely sodic and potassic rocks, often rich in carbonate. The carbonatite complexes of East Africa are considered to be the subvolcanic equivalents of such unusual volcanoes, which erupt sodium carbonate lavas and ash (Oldoinyo Lengai, in Tanzania).
- Post-rift, thermal subsidence is initiated by cooling of the asthenospheric material that was upwelled to lithospheric levels during rifting. The post-rift sediments are mostly marine, pelagic and poor in terrigenous material. The thermal subsidence seizes wide zones beyond the rift zone itself, but the syn-rift rugged morphology of the basin may be still reflected by various types of sediments. The relative elevations - pelagic swells are sediment-starved with typical condensed facies, sedimentary gaps, Neptunian dikes etc. The basins are marked by hemipelagic and eupelagic sedimentation, discontinuously affected by resedimentation events bringing material from nearby elevations.

Rifting modes

The stress field resulting from body forces and plate boundary forces within a plate may be locally modified by lithospheric or sublithospheric forces. For instance, the head of a buoyant mantle plume generates horizontal tensional stresses in the overlying lithosphere and can help, along with subcrustal erosion, to split the plate.
Two rifting modes have been envisioned, which refer to the role of the asthenosphere: active versus passive rifting. These two modes are distinguished in their initial stage, which the amount of associated magmatism discloses. These two modes are distinguished in their initial stage.



Mantle-activated, "active" rifting

The "mantle-activated" or "active" mechanism considers that hot and buoyant mantle plumes or diapirs initiate rifting. The upwelling asthenosphere bends the lithosphere about a large, dome-shaped topographic elevation up to several thousand km across (the Hoggar and Tibesti uplifts in central Sahara are modern examples) on top of which radial fractures make rifts. The lithosphere is thinned from below either thermally (heat input displaces isotherms upward) or/and mechanically (convective removal of lower lithosphere mantle) and rifts propagate down topographic gradients away from the upwelling dome. The main fractures finally form three main grabens that merge at the center, in a triple junction.
Plume-generated rifts begin with doming and extensive, pre-rift volcanism, predominantly alkaline, on continental crust and abundant igneous intrusions, commonly gabbroic or doleritic sills in the lower continental crust. Active rifts tend to be symmetric above the ascending asthenosphere and have wide and comparatively high shoulders; fault escarpments may range up to 2000 m in height and the total throw on these faults reaches 3-4 km in some places. The East-African rift has been considered to be typically mantle-activated.
Rapid rise of the plume causes adiabatic decompression and therefore partial melting of the mantle (up to 30%). Consequently, plume-generated rifts begin with extensive magmatism found as predominantly alkaline pre-rift volcanism and abundant igneous intrusions, commonly gabbroic or doleritic sills on and in the continental crust. Owing to the relatively shallow plume head and abundant magmatism, the high heat flow produces a very high geothermal gradient (up to 120°/km). The high thermal conditions reduce the density of the mantle. The isostatic response is a wide uplift and high-topographic region. For these reasons, active rifts tend to be symmetric above the ascending asthenosphere and have wide and comparatively high shoulders; fault escarpments may range up to 2000 m in height and the total throw on these faults reaches 3-4 km in some places. The East-African rift has been considered to be typically mantle-activated.

Lithosphere-activated, "passive" rifting

The alternative "lithosphere-activated" or "passive" mode attributes rifts to lithosphere extension under far-field tectonic forces. Within-plate tensile forces can be generated at the plate margins by slab pull (the slab pulls on the plate that also carries a continent, behind) or trench suction and tension is transmitted far into the plate interiors. Structural (pre-existing fault zones), thermal (high heat flow) and compositional (low-strength rocks) heterogeneities may stimulate deformation localization into particular zones.

Passive rifts begin with narrow grabens of clastic sedimentation and younger, limited volcanism. This rifting mode is termed passive because mantle upwelling is the submissive result of lithospheric thinning. Subsequent mantle decompression melting may produce the minor volcanism. In contrast to active rifts, the mechanical response of a stretched lithosphere tends to generates asymmetric systems involving a lithosphere-scale detachment fault. The magma-starved Baikal and Rhine grabens have been considered to be typically lithosphere-activated.

Geological documentation on rift zones shows that both modes exist and probably work simultaneously, with one or the other dominating. In addition, passive rifts may evolve into active rifts by the upward intrusion of the asthenosphere into the necked lithosphere.

Rift propagation

If there are several plume-generated graben systems not far from each other, their neighbouring fractures may connect into a continuous, but irregular rift zone. Propagating rifts gradually break through lithospheric plates. The orthogonal combination of across-rift extension and lengthwise propagation produces a characteristic V-shaped wedge of thinned lithosphere with a progressively younger age of extension structures in the propagation direction. One of the three main graben directions above a plume may become inactive to form a failed (aborted) rift arm. Alternatively, all three arms may develop into oceanic basins to install an intraoceanic triple junction (e.g. the junction of the Antarctic, South American and African plates). Two arms of the Afar triple junction of the Afro-Arabian rift system have already evolved into the Red Sea and Gulf of Aden floored by oceanic crust and it is expected that the third arm - the East African rift system will also develop into an oceanic channel in a couple of millions of years. Then it will split the African plate into two - the Nubian and Somalian plates.

Aulacogens: Failed rifts

A rift that did not lead to continental separation remains preserved within the continent as a failed rift or aulacogen (e.g. the Benue Trough of central Africa is a failed arm of the rift system that opened the Atlantic Ocean). If the lithosphere stops spreading at this stage, the hot mantle below the thinned lithosphere slowly cools and becomes denser. The now heavy mantle material acts like a sinker pulling down on the overlying crust, which causes regional subsidence. The bordering mountain ranges erode and supply sediments in the subsiding area. Isostatic adjustment to the weight of the sediments causes additional subsidence. Aulacogens are therefore often filled with thick sediments and contain ore deposits associated with their original activity.

Characteristics

Aulacogens are long-lived, narrow and elongate depressions that extend at high angles from the margins towards the interiors of cratons, yet remaining genetically related to larger oceanic basins into which they opened. The aulacogens actually are deeply subsiding, fault bounded sedimentary troughs first described in the USSR in the 1960s as “transverse basins and transverse fractures” that segmented the cratons. They exist in the geological record back to the oldest one, the 3 Ga-old Pongola in southeastern Africa. They have few characteristics:

- They are commonly located at re-entrants on continental margins.
- Their initiation is contemporaneous with continental rifting and associated with alkaline magmatism.
- They contain a much thicker sedimentary section than that found on the surrounding platform.

- They have a long duration, similar to that of adjacent active plate margin.
- Usually undeformed or gently folded, they tend to be reactivated by renewed faulting and subsidence (so-called posthumous movements), which eventually involves weak folding and metamorphism.

Origin

The concept considers that aulacogens are believed to result from the initiation of an RRR-triple junction within a continental plate, above a thermal plume. Three rifts meet above the thermal plume at angles of 120°, which is the least work and hence favoured configuration to release the doming-imposed stresses. As plate separation proceeds, two of the arms propagate and link up to form a single extensional plate boundary along which an ocean will open. The third arm becomes inactive and remains preserved in the continent as a failed rift. Alternatively, aulacogens could develop a RRF triple junctions but the case is rare. Therefore, depending on where and when failure occurred, either continental crust or oceanic crust can floor aulacogens. Other rare cases include different modes of continental extension forced by events or irregular impingement at plate boundaries.

Evolution

All three arms share the early tholeiitic basaltic phase, but usually only the failed arm develops as an alkaline or peralkaline eruptive province (e.g. the early Tertiary province of E Greenland).
Following initial rifting, they are very favourable locations for a river system carrying detritus from craton (e.g.: Niger River) with marginal fanglomerates, arkoses, mudstones, and locally playa evaporites. After opening of the ocean along two arms of the system, the aulacogen may localize continental drainage and nucleate large deltas at its mouth (e.g. Mississippi embayment).
If the ocean closes the aulacogen will be preserved at a high angle to the fold belt. Compression may induce axis-parallel yet weak folding (e.g. the Dnyepr-Donetz) or a new rifting phase, or strike slip faulting along the boundaries. These different eventualities are important for hydrocarbon exploration.

Continental rifts basically represent zones of extensional strain localization. The rift zone continuously widens while its margins move away. Once the continental crust is entirely thinned out, the rift becomes an oceanic basins and the extension zone becomes the constructive plate boundary. As such, rifts may represent the initial stages in the evolutionary cycle that further separates the older rifted segments and finally leads to continental break-up and ocean basin formation between two separated pieces of continental lithosphere.

Passive continental margins

The natural evolution of divergent plate boundary is recorded along the margin of an original continent. Extensional structures attest for rifting to some critical threshold of thinning, when the earlier continent was split and fragments dispersed by the sea-floor spreading process. The resulting plates consist of both old continental and new oceanic lithosphere while the new continental margins become passive intra-plate structures, meaning with nearly no seismic activity, as the spreading ridge moves away.

Formation

Stretching and thinning of the continental lithosphere at the margins of the split continents results from initial rifting followed by extensive and continued normal faulting. Rifting begins as a rather symmetrical and narrow extension zone between conjugate faults and involves lithospheric necking. Faults of the upper crust merge into or are cut by a flat detachment fault along the attenuated, ductile lower crust. The faults of one rift side evolve as the dominant break-away system. The continental crust then undergoes extreme extension in an asymmetric system of listric faults while the lithospheric mantle is thinned, possibly between two conjugate shear zones bounding the necking region. Sediments are generally deposited directly on tilted and eroded basement blocks, forming a profound unconformity. Synsedimentary faulting is common.

The distance from the earth's surface to the top of the asthenosphere is steadily reduced as stretching proceeds. Decompression melting produces volcanism into the rift zone. Alkaline igneous intrusions are common at that stage, with some of the magma being trapped at the base of the crust (magmatic underplating). Ultimately, the lithosphere is stretched and thinned to a point of rupture, permitting break-up of the continent. Magmatism evolves from alkaline to tholeiitic, with intrusion and extrusion of basaltic magma through the stretched lithosphere. The asthenosphere rises up to only several km below the ridge to form a magma chamber that continuously feeds the submarine volcanic edifices of mid-oceanic ridge basalts (MORB) that create a new oceanic lithosphere.

Morphology

The edge of the continent is a passive margin in the sense that it does not deform further (the reference examples are the Atlantic margins). Structures and rock associations of the rifting phase are frozen into rocks of the continental margin and they are passively conveyed laterally along with the rest of the plate of which the margin is part.

Main parts

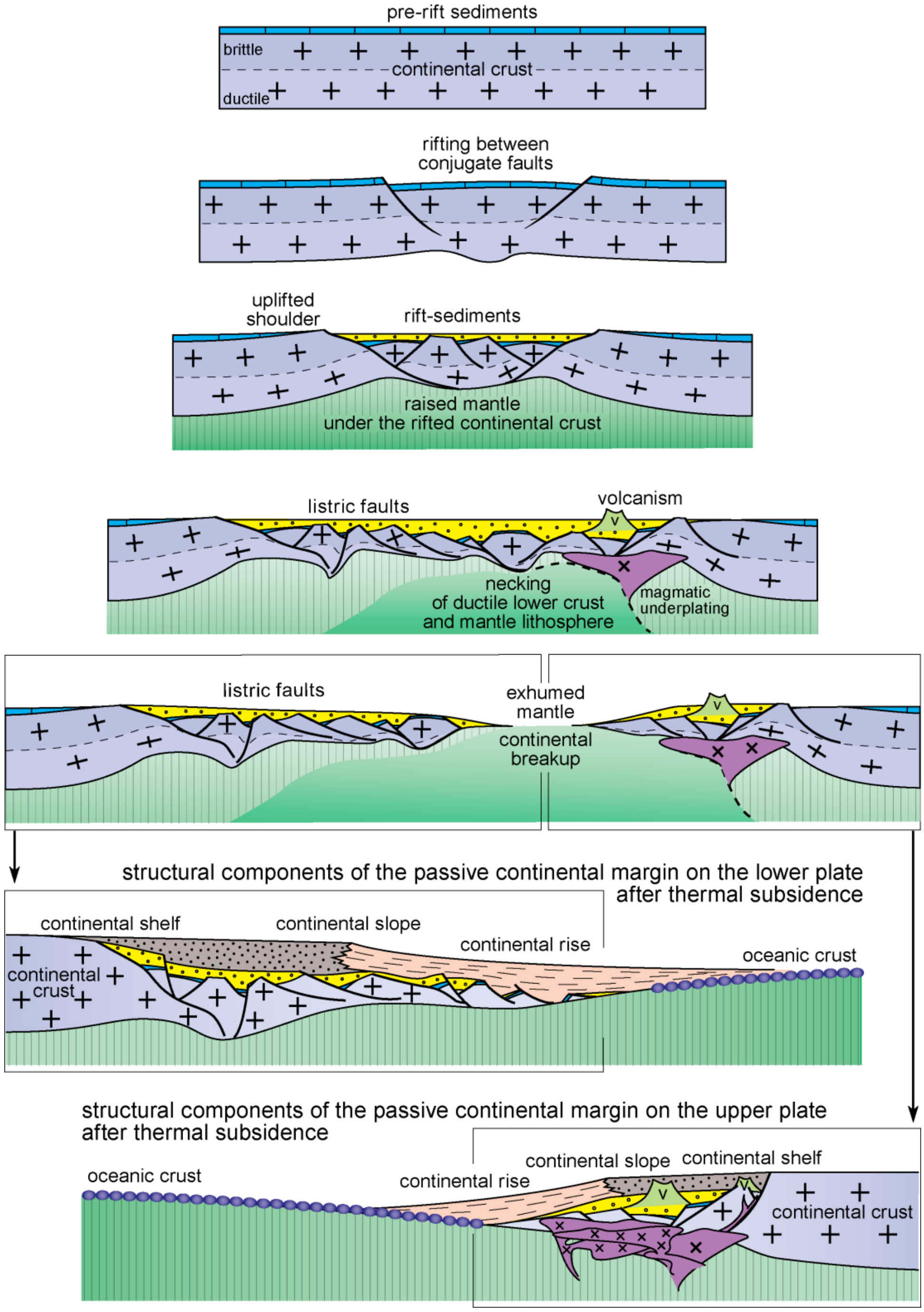
The oceanic crust is welded directly onto continental crust. The mixture of original continental rocks and the added oceanic component at the continental edge produces a hybrid transitional crust. Passive margins include three main parts:

- A coastal plain and a submarine continental shelf of variable width (from a few kilometres, e.g. Corsica, to over 1000 km in northwestern Europe). They are generally underlain by a thick sequence of shallow-water mature clastic or biogenic sediments. The shelf to slope transition is often reinforced by massive bodies of carbonate buildups, or reefs (e.g. the Great Barrier Reef of northeastern Australia) and the shelf basin behind is filled with thick shallow-water lagoonal carbonate and detrital sediments.
- The continental slope, at the edge of the continental shelf, which is generally present at the point where the shelf passes into a steeper topographic slope (3-5°) towards the basin. The slopes are unstable and deeply cut by submarine canyons conveying sedimentary material from the shelf to the continental foot and adjacent abyssal oceanic plains. The continental slopes are covered by unevenly thick aprons of sediments derived from the shelf and from the barrier reef. In front of great rivers, exceptionally thick and wide fans of clastic sediments are deposited in the river deltas (e.g. the Mississippi or Nile deltas). Most commonly, they are comprised of a sedimentary wedge that thickens seaward from a feather edge to 15 km or more.
- The continental rise which links the continental slope to the ocean basin. A relatively thick sequence of deep-water sediments is generally present along the continental slope and rise.

Upper-plate and lower-plate margins

After lithospheric necking abiding by the simple shear model, the master detachment fault soles a break-away system whose asymmetry is reflected at all scales in the margin structure and history. From the point (in two dimensions) where the major break-away fault reaches the bottom of the crust, the major detachment tends to follow the crust-mantle boundary, a surface of rheological contrast, before ramping down along the boundary of the upwelling asthenosphere. The general, shallow-dipping attitude of the major detachment fault (whose depth actually depends on rheological conditions) separates lithospheric-scale hanging walls and footwalls, which will be the two opposing, but structurally different passive margins after lithospheric breakup. The gross structure thus depends on the depth of decoupling between the hanging wall and footwall of the major detachment, hence on rheological layering of the extended lithosphere.

The hanging wall (upper plate) rifted margin is characterised by its comparatively thick continental crust with a narrow continental shelf and a thin sedimentary cover generally deposited directly, with a strong unconformity, on weakly tilted and eroded basement blocks. The early rift structures with syn-rift sediments (including evaporites) are preserved on the continent-side of the margin. The major detachment separates the crust from its mantle (delamination) and normal faulting across the mantle causes stretching and rupture of the lithospheric mantle before or at the same time as crustal separation. The asthenosphere raises (upwells) to replace the stretched/thinned lithospheric mantle gone with the footwall plate. Delamination leaves rafts of continental crust below which the lower crust has been attenuated or removed along with the mantle lithosphere. Upwelling generates adiabatic decompression, hence melting of the lower mantle and of the asthenosphere. The resulting magma is added to the volcanism of the initial rift, which has been shifted away from the break-away fault system with the hanging wall. Consequently, upper plate margins are characterized by voluminous magmatism over a short time span (Atlantic margins of Brazil and Greenland). Accordingly, they are also termed volcanic rifted margins.



The footwall (lower plate) rifted margins represent the footwall shoulders of early rifts with few remnants of syn-rift sediments. Lower plate rifted margins consist of abruptly thinned continental crust with a wide shelf and a thick sedimentary cover on highly rotated blocks displaying exhumed middle- to lower crust at their base. Breakup and stretching of the entire pre-rift crust results in the denudation and exposure at the seafloor of the continental lithospheric mantle across a several tens of kilometres wide continent-ocean transition zone, where the mantle was the direct footwall of the major detachment. Consequently, wide areas of serpentinized peridotite characterize lower plate margins. In the transition zone, thin allochthonous blocks of continental crust may directly overlie the exhumed mantle serpentinites, which make the seafloor. Isostatic compensation after thinning produces more post-rift, thermal subsidence than in upper plate margins. There is nearly no magmatism associated to such margins (Atlantic coasts of North America, northwest Africa, West Iberia). Accordingly, they are also termed magma-poor rifted margins.

Thermal subsidence

Thinning produces heat advection, thus intensifies geothermal gradients as magma and deep crustal rocks carry upward the heat attached to them. The thermal input is so great during initial stages of rifting that the margins of the rifted continents are buoyantly uplifted. With sustained seafloor-spreading, the two continental margins move progressively farther from the hot mid-oceanic ridge and can cool down. Colder rocks are denser, so the edges of the previously thinned continental lithospheres gradually subside below sea level to maintain isostatic equilibrium; a non-mechanical process termed thermal subsidence. This subsidence opens space for deposition of post-rift sediments. A pericontinental sea forms on the continental shelf. Shallow marine conditions may spread far over the continent to form an epicontinental sea (e.g. North Sea). Massive subsidence (up to 10-15 km) of the passive continental margin takes place as the attached ophiolitic rocks cool off. Most sediment on Earth lay on passive margins which, therefore, contain more than half the world’s oil reserves. The total post-rift subsidence is related to both thermal down-warping and sedimentary load. However, sea-level fluctuations exert a strong control on the accommodation history and, consequently, the stratigraphic framework of passive-margin basins.

Note:
Isostatic calculations suggest that, in addition to thermal subsidence, the weight of sediments pushes down the basement by about a third of their thickness

Exercise
Calculate how long it takes to recover a normal geothermal gradient after lithospheric thinning.

Ridges: plate divergence in oceanic setting

The transition to the oceanic stage, the "continental break-up," is marked by the break-up unconformity, between tilted and faulted sediments below, and undeformed sediments, above. This unconformity indicates when stretching between the two diverging plates stopped to deform the two conjugate continental margins and was taken up by the generation of the intervening and new oceanic lithosphere.

Morphology

The present day divergent plate boundaries reside mainly in oceans where they form broad, fractured swells. The prominent physiographic expression is a world-encircling, ca. 100 km wide and symmetrical topographic relief that rises up to 3000 m higher than the adjacent ocean floor. It is the ocean ridge system, which wraps around the globe like an approximately 70 000km long seam of a tennis ball. Almost all fault plane solutions from ridge crest areas show normal faulting. The gravitational field over these submarine mountain ranges shows that they are isostatically compensated but do not have deep crustal roots. These mountains are high because they are underlain by young lithosphere, which is hot and, therefore, less dense than the older and colder adjacent oceanic lithosphere. Being cut off from terrigenous supplies, the mid-oceanic ridges are sediment-starved. The thickness of the basaltic crust is influenced by spreading velocity, temperature and composition of the upwelling mantle.

The Structure of the mid-oceanic ridges depends on the spreading rate. Slow spreading ridges (< 4 cm.a-1, e.g. the Mid-Atlantic Ridge) are narrower and have a more pronounced topography than fast-spreading ridges (> 4 cm.a-1, e.g. the East Pacific Ridge) that lack a typical rift valley. Instead, elevation of the seafloor of several hundred metres forms an axial high. This morphological difference may reflect differences in rates of magma supply.

Spreading dynamics

The dynamically active part of the system is restricted to a prominent axial rift valley that actually marks the plate boundary. Within the 1-2 km deep, 5-30 km wide rift zone, the opening between diverging plates is continuously filled in by olivine tholeiite magma issued from decompression partial melting of the underlying and upwelling mantle peridotites. New oceanic lithosphere is created by the combination of intrusion of mafic igneous rocks, extrusion of tholeiitic basalts interlayered with oceanic sediments, and extensional faulting. A spatial variation between high-Mg and low-Mg basalts may reflect the different rates of sea-floor spreading between different segments of the same ridge. Off-axis volcanism, whose relationship to the magmatism at the axis is obscure, produces a high-Mg basalt suite that may include basaltic komatiites and picrites. When cooled and crystallised, the intrusions and freshly accumulated Mid-Ocean-Ridge-Basalts (commonly called MORB) with their thin sedimentary cover become part of the moving plates and thus constitute new additions to the lithosphere. Accordingly, plate boundaries along oceanic ridges are also called constructive boundaries. As new lithosphere forms, it continually spreads away from the ridge at rates of several centimetres per year. This process is termed accretion.

Plates continue to cool as they move away from the ridge, and they thicken and become denser in the direction of motion; as a result the lithosphere subsides and there is a direct relationship between water depth and age of the lithosphere. The characteristics of a mid-oceanic ridge depend on its spreading rate. Slow ridges (e.g. the Atlantic) are higher and more rugged topography than the fast ones (e.g. the East Pacific Rise). The topographic elevation of the ridge is due to the greater buoyancy of the thinner, hotter lithosphere near the ridge.

As lava cools in the rift, it acquires magnetization. Each newly formed element of oceanic crust acquires the direction of magnetization of the prevailing magnetic field. During spreading, magnetized seafloor moves laterally and symmetrically away from the ridge, which results in the magnetic anomaly stripes used to calculate plate movements and their rates over the last 200 Myrs.

In response to changing tectonic conditions, a ridge may grow or propagate into an adjacent plate. Whole sections of a ridge may “jump” to form a new rift parallel to the existing ridge. Importantly, if the rate of seafloor spreading accelerates, the ridge volume increases and displaces seawater, which results in a global increase in sea level. Variations in the rates of seafloor spreading are the primary causes for worldwide changes in sea level.

Ridge segmentation

The ridge axis undulates up and down in a systematic way, defining a fundamental partitioning of the ridge into segments bounded by a variety of discontinuities, principally transform faults orthogonal to the ridge axis. Wavelength and amplitude are larger along fast than along slow spreading centers. These variations may reflect enhanced magma supply along shallow portions and starved magma supply at segment ends, where transform faults and their fracture zones usually perpendicular to the ridge axis offset all known mid-ocean ridges. The ridge activity is also affected by the juxtaposition of thick cold lithosphere against the end of a spreading segment.

Extension at convergent plate boundaries

It may seem paradoxical that extension occurs in converging systems. Back-arc basins are common around the Pacific (Aleutian Basin, Okhotsk Sea, Japan Sea, Ryukyu Trough, South China Sea) and are floored by the oceanic lithosphere generated along irregular systems of active and extinct, even

crosscutting spreading ridges. They indicate that extension commonly occurs above subduction zones. Marginal basins are rather short-lived, mostly inactive back-arc basins at present. They may have formed transiently under locally favourable circumstances such as differences in movement vectors and velocities of the upper and lower plate, or changes in the slab inclination. Back-arc extension seems to be controlled by the dip of the subducting slab.
Low-angle subduction generates compression in the upper plate, since the subduction fault is longer in its contact with the upper plate. Therefore it is subjected to frictional slipping and the causative slab-pull is transmitted as a collisional resistance on the upper plate generating compressive stresses. The South American Andean margin is a typical example.
Steeply dipping slabs cause small-scale convective flow of the mantle beneath the arc, but above the downgoing slab: the slab may drag the viscous mantle with it, causing hot asthenosphere from deeper in the mantle to flow upward and take its place. The resulting convection pattern thins the overriding lithosphere and may exert tension on this upper plate lithosphere. Extension may result in a new sea floor spreading in or behind the island arc (Philippine Sea behind the Mariana Arc). The magma-generating process is located in the upper mantle and produces tholeiitic basalts similar to those seen in the oceans, although variations include the presence of a high-Al group of basalts produced in the early stages of back-arc rifting. The floor of the back-arc basin is younger than the subducting oceanic lithosphere and causes oceanward migration of the arc and its forearc with respect to the continent.
Furthermore, steep slabs tend to move away from the trench in the direction opposite to plate movement, thus developing extensional structures in the fore-arc region. Subduction retreat (rollback) also generates trench suction forces that exert extensional stresses on the upper plate margin. Rifting and extension of the upper plate is enhanced by its thermal weakening caused by mantle up-welling and volcanism above the slab. The retreating slab hence controls in which direction the overriding plate is allowed to extend.
Spreading may separate a piece of the arc from the active arc. A basin is then opened between the extinct remnant arc and the active arc. Depending on proximity to the arc, sediments are volcanoclastic, hemipelagic or pelagic. One important variable that may control compression versus extension in converging systems might be the angle of the subducting plate.

Ridge-associated triple junctions

A plate boundary ends against another plate boundary; the intersection is a triple junction. Six major ridges dominate the present-day Earth. They terminate either against transform faults, trenches or at triple junctions. Three ridges (RRR) triple junctions form the most stable type as in Rodriguez and Galapagos. Other terminations include RRT (East Pacific Rise and Southeastern Indian Ocean), RTT (none presently active), RFF (Owen Fracture- Carlsberg Ridge) and RFT (southern end of the San Andrea Fault) junctions. RRF junctions are one of the two unstable modes.

Balanced cross sections

Accurate geological cross sections are of paramount importance when large commercial investments are at stake. For this reason balanced section techniques have concentrated much effort from structural geologists.

Concept

The idea is that one must be able to restore the structural interpretation to its original geometry and to its correct palinspastic reconstruction while the volume of rocks is kept constant throughout deformation. The balanced cross-section is restorable, retro-deformable; it is admissible (i.e. it represents the structure of the region) and viable (i.e. its pre-deformation state can be reconstructed without geometrical gaps or overlaps).
The basic approaches assume plane strain, i.e. the structural development takes place entirely within the section plane; no material is entering or leaving the plane of section. This implies preservation of areas from the initial to the final state in the section plane. Volume (area in 2D) preservation confronts

several and frequent exceptions such as dissolution, erosion, mobile layers (salt in particular), etc. In cases there are incompetent, mobile layers, at least one reference bed is assumed to be strong enough to retain its original length and thickness. Any overlap or void in the restored stratigraphic section points out a volume imbalance where interpretation or correlation is incoherent. In that sense, balancing cross sections is an important test of the validity of the structural interpretation.

Section balancing is applicable to basin infill by relating hanging wall shapes to fault geometry, with the prime assumption that the footwall is rigid below the detachment horizon. In addition to better understanding the geometry of the studied structures and the relative role of different layers, balancing helps determining the sequence of faulting events, hence provides a kinematic evolution of the system.

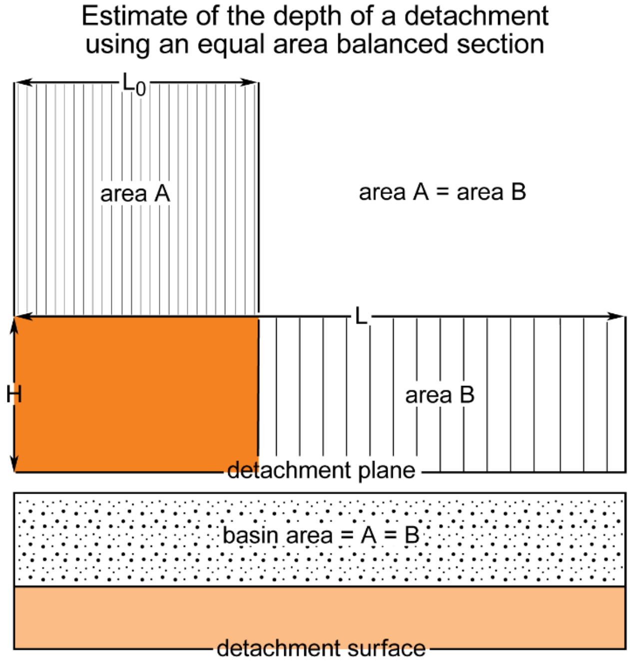
Balancing sections of sedimentary basins further implies that the compaction behaviour is known and that time-depth conversion through decompaction of sediments is correct. In the low temperature, brittle deformation regime of upper-crustal basins, internal strain of rocks is negligible. Thus the 3D volume consideration can be reduced to 2D plane strain. Flow of shales and salt diapirs are exceptions to this rule. Different techniques make different assumptions about how the hanging wall deforms with respect to the fault geometry; this also implies that the footwall does not deform. But all assume that the transport direction is subperpendicular to fault strike. Cross sections must then be balanced parallel to the movement direction, whether there is 3D-volume / 2D-area balance or bed length balance.

Area-balanced construction

This technique is used to predict the depth H of the flat part of a listric fault. The premise is that if the rollover shape is known, then the fault shape can be predicted, assuming plane and isovolumetric strain (or taking decompaction into account). The profile area of the related basin (the A_B area dropped below the initially flat reference horizon) is equal to the rectangular profile area of hanging wall moved out of the pre-extension system boundaries above the detachment surface. Taking L_0 the initial length of the system (footwall + hanging wall bed length) and L the length after extension:

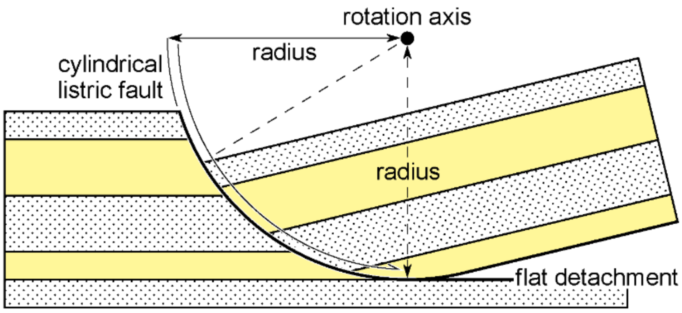
$$A_B = \left(L - L_0\right) H$$

This is in any case a rough estimate since deformation within the hanging wall may result not in a rectangle but a trapezium or other shapes, for example because of non-constant bed length (layer-parallel strain). Then the equation must be modified accordingly.



Cylindrical listric fault

Assuming the fault profile follows a circular arc, the rotation geometry can be used to model depth to detachment: The rotation axis is defined by the intersection of two lines orthogonal to the fault plane. The radius defines the distance from the rotation axis to the flat detachment.



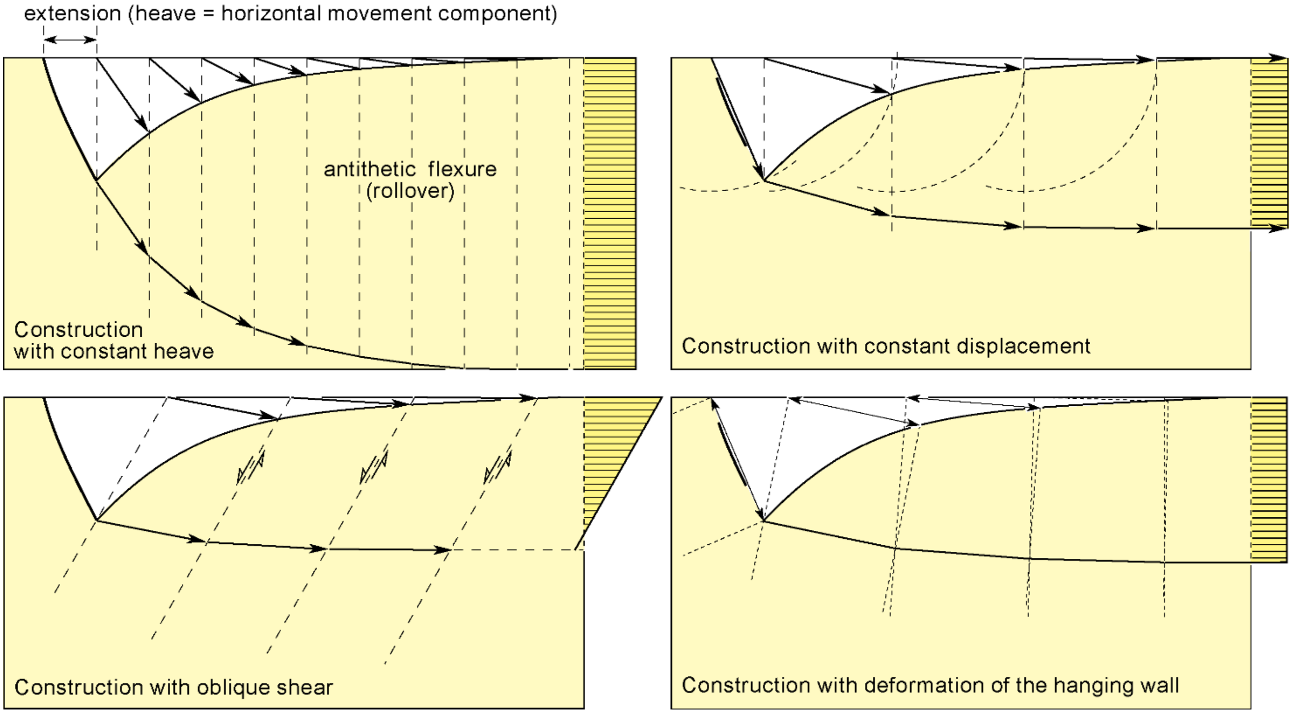
Model of cylindrical, listric normal-fault to infer the depth of detachment

Vertical shear (constant heave) construction

This technique is used to determine the shape of the listric fault from the rollover geometry. It assumes that horizontal translation of the hanging wall is constant and equal to the magnitude of extension. The hanging wall deformation is accomplished by vertical simple shear, which represents collapse along vertical slip planes to fill the void between the footwall and hanging wall. The technique consists in dividing the displacement vector along the fault plane into heave (horizontal component) and throw (vertical component). The heave is considered to be constant all along the length of the fault. If there are no synsedimentary faults, then the curvature of the rollover flexure has a direct geometrical relationship with the form of the listric fault. The procedure is as follows:

- Define on a geological (or seismic) profile of the footwall a stratigraphic horizon that is bent in the hanging wall rollover.
- Measure the heave from the horizontal stratigraphic separation of the reference layer.
- Divide the hanging wall cross section into vertical segments with width equal to heave. These vertical lines intersect the upper horizontal surface and the rollover surface at successive points.
- Draw vectors from the intersection of one of these vertical lines and the horizontal surface to the intersection of the next line with the top of the rollover flexure. These vectors gradually become shallower and shorter away from the breakaway into the hanging wall.
- Translate downward, vertically, each vector of each vertical segment to add it to the previous vector, (place tail at previous head of arrows with respect to the regional transport direction, from the fault breakaway). The gradually flattening out fault plane is constructed with its segmented displacement by the translated vectors.

With this technique, the bed thickness along the vertical shear planes stays constant and the hanging wall is area balanced. However, the orthogonal thickness of the strata is variable and the bed length increases.



Four geometric methods
to reconstruct the shape of a listric normal fault from the same shape of the roll-over antiform

Constant slip construction

The vertical shear (constant heave) construction can be adapted to different prejudices. For example, it is modified if the displacement vector along the fault plane is taken constant rather than the heave. Then the width of the successive vertical segments is defined as follows:

- Define the movement vector on a profile and draw the first vertical line.
- From its intersection with the horizontal surface draw a circle with a radius equal to the slip vector, and draw a vertical line from the intersection of the circle with the top of the rollover flexure.
- Then proceed successively from each new surface/vertical line intersection to obtain the next point and draw from point to point vectors to the top of the rollover flexure.
- Translate downward and add vectors to determine the fault geometry in the same way as in the constant heave construction.

Inclined shear constructions

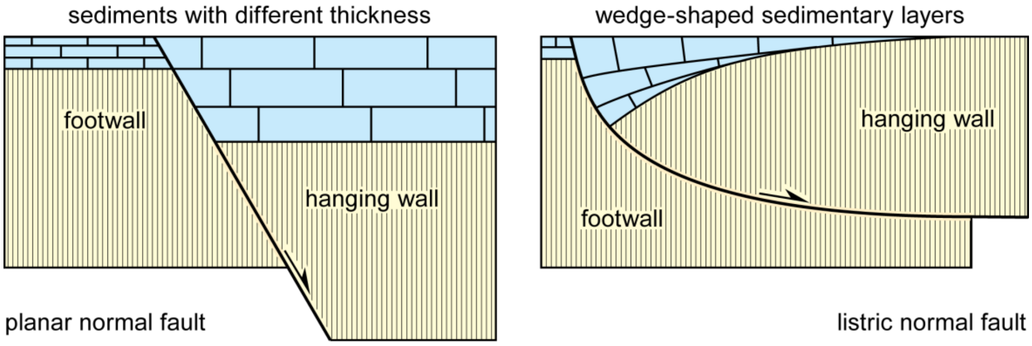
Additional adaptations include:

- Inclined simple shear or simple shear along planes parallel to synthetic and antithetic faults that obey Coulomb failure criteria (for example, conjugate faults dipping 60°, i.e.at 30° to the vertical maximum principal stress);
- Variable shear directions;
- Hanging wall deformation through material movement parallel to the fault profile;
- A combination of vertical simple shear and constant bed length, which requires flexural slip folding to generate the roll-over structure;
- Compaction in the hanging-wall.

All these techniques yield different geometries and all of them give a rough estimate of the true geometry. The assumption that volume (area) remains constant is a weighty decision that often violates natural deformation. As a consequence, section balancing is a valid exercise but one should remember that it may lead to geometries that are restorable but flawed.

Normal faulting and sedimentation

Large-scale extension is generally associated with basins and synchronous sedimentation. Synsedimentary faults (growth faults) move while grabens and half-grabens are progressively filled with sediments. Growth faults are therefore recognised by an abrupt change in thickness of beds, thicker on the downthrown hanging wall

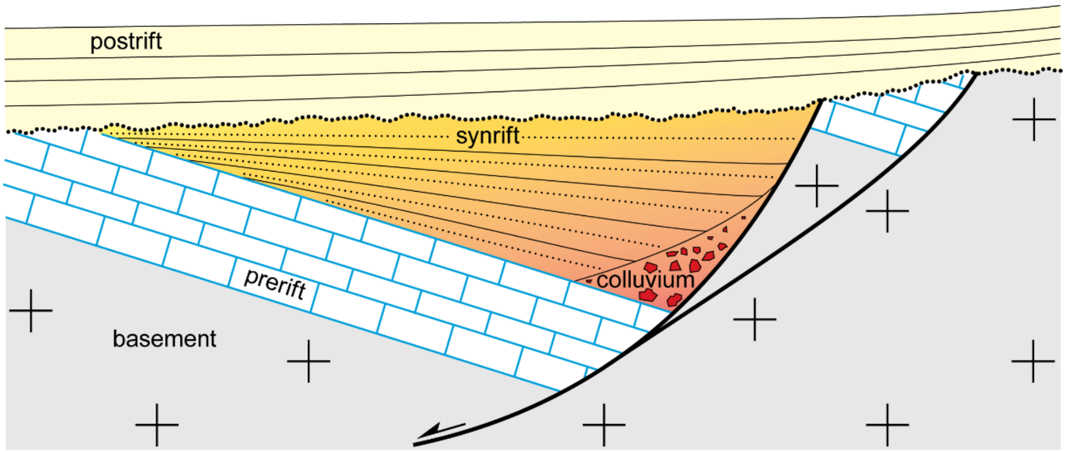


Growth normal faults with associated sedimentary basins

Infill of a half-graben typically decreases away from the major listric fault, thus defining sedimentary wedges. Dip and compaction increase downwards, i.e. with age of sediments. As a consequence, the displacement and the amount of rotation decrease upward, from older to younger beds.

The time relationships between faulting and sedimentation led to a three-fold classification of sedimentary sequences in passive margins:

- Pre-rift sequences were deposited before rifting, in a tectonically quiet environment. The series of parallel layers is attached to the basement.
- Syn-rift sequences were deposited during rifting. Since most normal faults are listric, block rotation during faulting produces sedimentary wedges with onlaps and soft-sediment deformation (e.g. slumps). Colluvial wedges may exist against the fault surface (old scarp).
- Post-rift sequences were deposited after rifting, during thermal cooling. They are parallel and sub- horizontal layers unconformable on all previous sequences.



Sedimentary sequences classified with respect to rift-faulting

Pangea break-up

Tectonic reconstructions of late Paleozoic times display one ocean, Panthalassa surrounding one supercontinent, Pangea.

Tethys

The break-up of Pangea began during the late Permian with continental rifting along the future margins of two continents, Gondwana to the south and Laurasia to the north of an east-west trending, equatorial oceanic basin, the Tethys.
Generalized rifting and subsequent ocean opening prograded westwards:

- Between Africa and Europe in Permo-Triassic times;
- Between Africa and North America during the Middle–Late Triassic;
- Between North and South America during the Middle Jurassic.

Continental drift accompanied sinistral wrenching between Laurasia and Gondwana from the Middle Triassic to the Late Jurassic.
Closure of the Mesozoic Tethys started during the Late Cretaceous and continental collisions began to occur during the Eocene, with Tethys remnants forming discontinuous ophiolites.

Present day oceans

Younger oceanic branches formed later to lead to the present-day tectonic configuration.

Gondwana break-up

Gondwana began to break up in the Middle Jurassic (ca. 165 Ma) when an eastern continent comprising Antarctica, Arabia, Australia, India, New-Zealand and Madagascar separated from Africa.
The South Atlantic Ocean started to open, along with the Benoue Aulacogen in Chad, during the Early Cretaceous (ca. 130 Ma).
New-Zealand separated from Antarctica between 130 and 85 Ma.
Australia separated from Antarctica in the late Cretaceous (ca. 80 Ma).
Madagascar and the narrow microcontinent of the Seychelles Islands were broken off India at the Cretaceous-Tertiary boundary (ca 60 Ma). The India-Madagascar-Seychelles separations coincide with the eruption of the Deccan Trap basalts, whose eruption site may survive as the Réunion hotspot in the Indian Ocean.
Australia and New-Guinea separated ca. 55 Ma.
Antarctica separated from South America during the Oligocene (ca. 30 Ma).
The active Red Sea and East African Rift show that the dismemberment of Gondwana is continuing.

Laurasia break-up

Around 180 million years ago, the North Atlantic Ocean began separating Laurentia (North America) from Eurasia.

Conclusions

At ca. 1100 Ma Rodinia was a supercontinent; it broke apart at around 800-700 Ma. Similarly, the break-up history of Pangea exemplifies dispersal of supercontinents. Extensional processes precede dispersal, resulting in crustal thinning and the emplacement of bimodal magmatism before rifts form and eventually evolve into small ocean basins such as the Red Sea. Continental rift tectonics arises from extension of the lithosphere, driven by intra-plate stresses or hot spots, and due to indentation tectonics associated with continental collision. The continental stage of rifting is accompanied by a typical three-fold succession of continental-alluvial red beds, evaporites, and marine carbonates. Once a small ocean basin has formed, sedimentary debris is shed into the ocean and a passive continental margin evolves with deposition of clastic and carbonate sequences onto the older continental crust.
Cessation of extension leads to development of an aulacogen. Continued extension may lead to continental breakup and formation of new oceanic lithosphere separating subsided continental margins. Continental margins are also subdivided into volcanic (upper plate) or non-volcanic (lower plate) margins, as a function of the amount of associated intrusive and extrusive magmatism.

Typically, normal faults in an extension system compose relay or parallel arrays of fault segments. Fault segments usually bend at all scales, both in map view and in cross-section. Movement on both planar and listric faults generally results in rotation of hanging wall blocks around horizontal axes, thus producing tilting of overlying blocks and/or formation of rollover anticlines and synclines.
Extensional strain localization depends on the rheological characteristics of the lithosphere, on boundary and initial conditions such as the applied strain rate. The style of deformation is mainly controlled by the competition between the total resistance of the lithosphere and the gravitational forces. An understanding of isostasy is critical to understanding the evolution of sedimentary basins. In particular, syn-extension and post-extension subsidence have critical consequences on the sedimentary environment. In interpretations and profile balancing, the choice of the correct model for the correct circumstances (i.e. pure shear versus simple shear models) is critical.

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