THRUST SYSTEMS

Thrust systems are zones where plates or crustal blocks move toward one another. Convergence may occur:

Between two continental lithospheres	e.g. Alps-Himalaya belt
Between two oceanic plates	e.g. Mariana-Philippine, Caribbean Islands
Between an oceanic and a continental lithosphere	e.g. Andes, North American Cordillera

There are four mechanisms for accommodating tectonic convergence:

- subduction
- volumetric shortening with localized thickening,
- lateral extrusion, and
- buckle folding.

Plates in convergence are in constant competition for space and the response to the space problem is very sensitive to the nature of converging lithospheres. Oceanic plates avoid confrontation and plunge into the asthenosphere (**subduction**) or climb onto continents (**obduction**). In contrast, continents **collide**, causing big damages such as mountain systems. In any case, horizontal transport and shear on shallow dipping thrusts predominates over vertical movements and the bulk result is crustal shortening and thickening taken up by compression structures. This is why, since the recognition of current subduction zones and their related features, plate tectonic concepts have been extensively employed to explain past orogen patterns.

Major tectonic structures around the Mediterranean Sea
adapted after Tapponnier P. (1977) *Bull. Soc. géol. Fr.* **19**(3), 437- 460

Compression structures (folds, thrusts) occur on all scales, from millimetres to kilometres, and develop at any crustal level, therefore under various conditions. A thrust system is an interconnected network of thrust faults that are usually also kinematically linked. Horizontal shortening forces the topography upwards, creating a mountain defined as a landform higher than the neighbouring area. Mountain building is a complex process, termed **orogeny**. The presence of mountains as physiographic features of **orogenic belts** (or simply **orogens)** is not integral expression of an orogeny. Erosion levelled ancient orogens to flatlands in the relatively inactive interior of continents. In addition, the structurally challenging parts of recent and growing orogens may not lie in the visible mountains; instead, they may be 10 or even 100 km below the Earth's surface.

The geometrical definition of compression structures, formed where the dominant stress is compression, is valid for the three stages of plate convergence defined according to their timing relative to ocean closure:

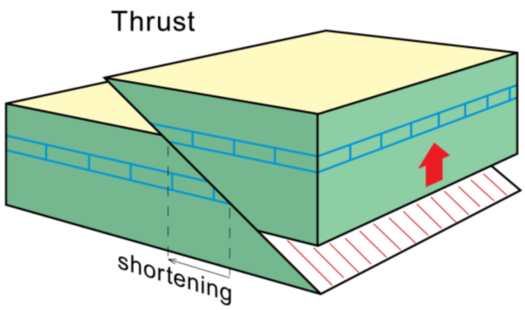
- Pre-collision tectonics involves subduction of oceanic material with the eventual formation of accretionary wedges and local obduction of ophiolites. Convergence causes thrusting of the more buoyant lithosphere over the less buoyant plate.
- Collision tectonics involves thickening and imbrication of the crust and lithospheric mantle when the ocean closes. Hence an orogenic belt is generally aligned along a zone of continental collision. Deformation produces excess topography (mountain range) that erosion modifies and destroys on a long term.
- Post-collision tectonics involves within-continent deformation effects of continued convergence after closure of the ocean. In particular, it includes gravitational instabilities formed by the thickened lithosphere. Even though most deformation is concentrated in boundary regions between plates, some regional structures form within the interior of plates, through transmission of tectonic stresses for great distances from plate boundaries. This is the case in Asia where the India - Asia collision has resulted in a very wide belt of complex structures on the Asian side of the suture. Intraplate deformation challenges our perceptions of mountains building and our understanding of stress propagation.

GEOMETRIC RULES OF THRUST-FAULTING

A lot of work has explored the geometry and kinematic of compression zones. Research in many thrust belts and related analogue and numerical modelling have revealed several recurrent characteristics that have led to the development of empirical, but not absolute rules regarding thrust geometry and growth. These few basic rules are valid only if the thrust area was not deformed (i.e. folded) before the considered thrust event.

Thrust faults - Basic terminology

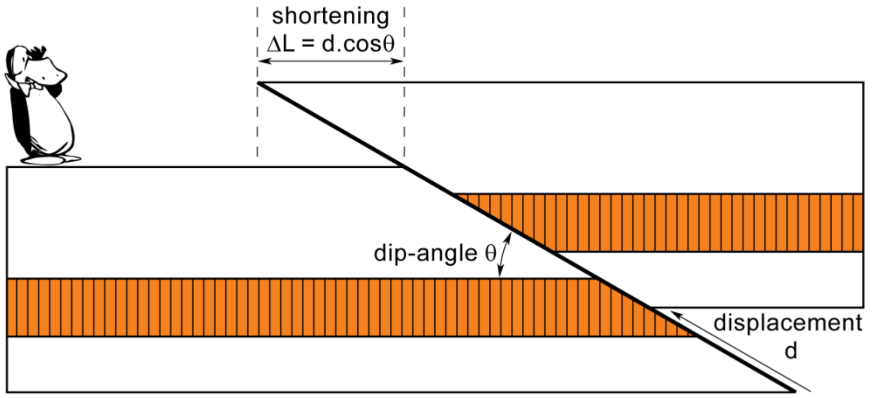
A thrust is a contractional fault that accommodates horizontal shortening of a datum surface, normally bedding in upper crustal rocks or a regional foliation surface in more highly metamorphosed rocks. Generally, a thrust places older strata over younger strata so that the stratigraphic sequence is generally **duplicated**.



Definition

In the French and in the German literature, a **reverse fault** occurs primarily across lithological units, therefore dipping > 45° in flat sedimentary regions where a **thrust fault** is a gently sloping (dip <<

45°) reverse fault. **Décollement** describes a thrust fault *within* or at a low angle to lithological units. Low-angle and high-angle thrusts may be different segments along the same fault surface because thrust faults are rarely planar; they are often listric (concave upward) and antilistric (concave downward).
Due to the thrust dip θ, shortening ΔL is smaller than the displacement d of a planar fault. The relationship is $\Delta L = d \cos\theta$.



Relationship between displacement and shortening on a thrust fault

An empirical relationship has been established between displacement (d) and width (W) of isolated thrust faults as being approximately:

$$d = a.W^{1.4}$$

where a is a constant.
The thrust displacement generally decreases up-dip so that a shallow-dipping thrust fault may terminates before it reaches the Earth surface. Thrusts that do not break the surface are called **blind thrusts**. Thrusts that reach the surface are called **emergent thrusts**. The termination of a fault at the surface of the Earth is the **fault trace**. Typically, superficial material is caught up and overridden by the advancing emergent thrusts.

Footwall, hanging-wall

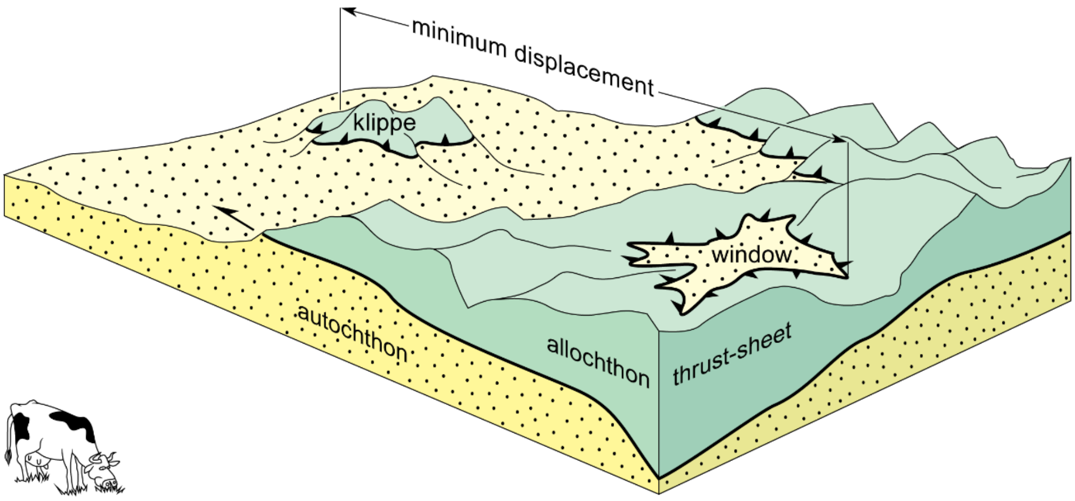
The rock immediately above and below a non-vertical fault or shear zone is the **hanging-wall** and the **footwall** of the fault, respectively.

Allochthonous / par-autochthonous / autochthonous

Overthrusting involves the displacement and tectonic emplacement of hanging-wall rocks forming **thrust-sheets (nappes)**. Rocks within thrust sheets have been translated great distances away from their original site; they are **allochthonous**. Allochthonous units often consist of subordinate thrust sheets that possess a common displacement history. They come to rest on **autochthonous** rocks, which have retained their original location, or on **par-autochthonous** footwall material if it has been moved close to its original location.
Thrust sheets **decoupled** from the underlying rock units by *décollements* tend to be thin compared to their horizontal dimensions and commonly exhibit a wedge shape, thinning from rear to front in cross section.

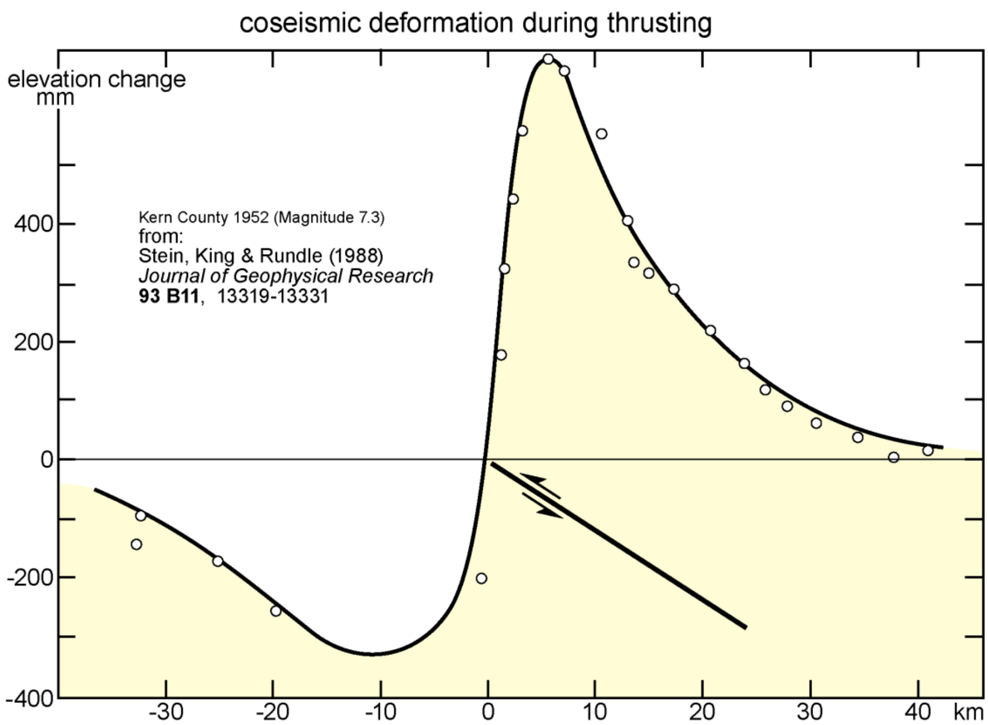
Erosion exposures: window and klippe

A **window** (or **fenster**) is produced when erosion made a hole through a thrust-sheet to expose the footwall rocks beneath the thrust fault; autochthonous or para-autochthonous rocks are completely surrounded in map view by rocks of the allochthonous hanging-wall.
A **klippe** is an isolated, erosion remnant of a thrust sheet completely surrounded in map view by rocks of the footwall.
Both klippen and windows are indicators of minimum displacement.



Surface deformation

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



Styles of thrust deformation

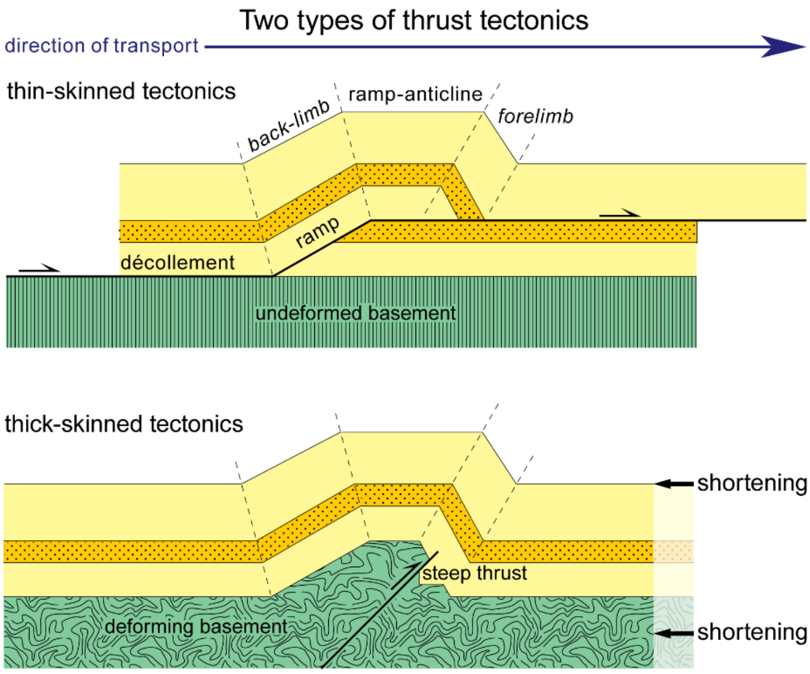
Foreland/hinterland

The **foreland** is the footwall area in front of the thrusts, towards which the thrust sheets moved. Forelands are the margins of an orogenic belt and are low topography regions in active mountain belts. Foreland sediments thicken towards the thrust belt. The region behind the thrusts is the **hinterland**, which defines the axial (interior) region of an orogenic belt. Hinterlands are regions of high topography and strong relief in active mountain belts. Thrust propagation proceeds from hinterland towards foreland.

Two deformation styles are commonly invoked to describe thrust tectonics and even put forward contrary models: **thin-** and **thick-skinned tectonics**. These two styles are rather characteristic of forelands and hinterlands, respectively.

Thin-skinned tectonics

Thin-skinned tectonics principally refers to thrusting that does not affect the basement whereas subparallel sets of folds and faults deform the cover. The initially subhorizontal sedimentary sequence is detached along weak **décollement horizons** (e.g. salt, shale, overpressured layers) and is deformed independently from the underlying substratum. Typically, deformation is confined to the cover while the basement slides underneath rigidly (no thrust cuts through).



Thick-skinned tectonics

In hinterlands, often the crystalline core-axes of mountain belts, deformation is principally controlled by high-angle thrusts and their interaction with the deforming ductile basement. Thrusting that involves basement deformation is termed **thick-skinned tectonics.** Combined basement-cover thrust sheets are designated as **basement-cored nappes**.

Thrust trajectory

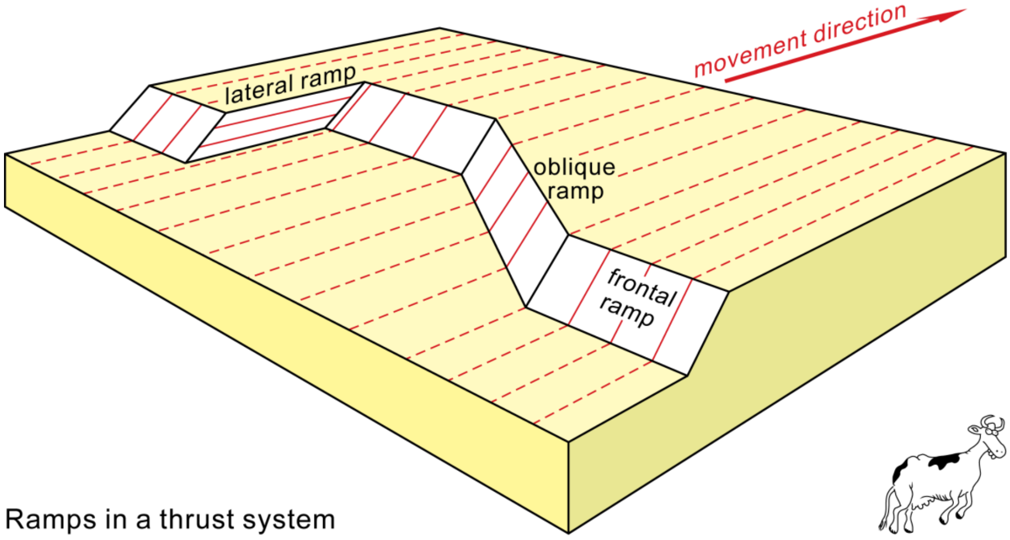
The thrust trajectory is the path that a thrust surface takes across the stratigraphy. Where thrusts involved in thin-skinned tectonics affect a set of nearly horizontal bedded rocks, they generally follow a **staircase trajectory** made up of alternating **flats** and **ramps**.

Flat

A flat occurs where a fault lies within, and remains parallel to a specific, typically incompetent stratigraphic horizon for a great distance. Flats, where the hanging-wall slides along a relatively weak bedding plane, are also called **décollement planes.**
Two parallel flats are distinguished as **floor** (bottom, sole) and **roof** (top) **thrusts**.

Ramps

Thrust-ramps occur where a fault climbs through a competent stratigraphic sequence, usually over short distances and typically at angles of 30-45° to bedding. Most commonly, thrust faults ramp up section in the direction of tectonic transport. **Frontal** ramps approximately strike perpendicular to the transport direction. In thrust systems, rocks are pushed together against the frontal ramps, which therefore are contractional. Ramps are also found **oblique** or parallel (effectively strike-slip) to the transport direction (**lateral ramp, transfer** and **tear faults**).

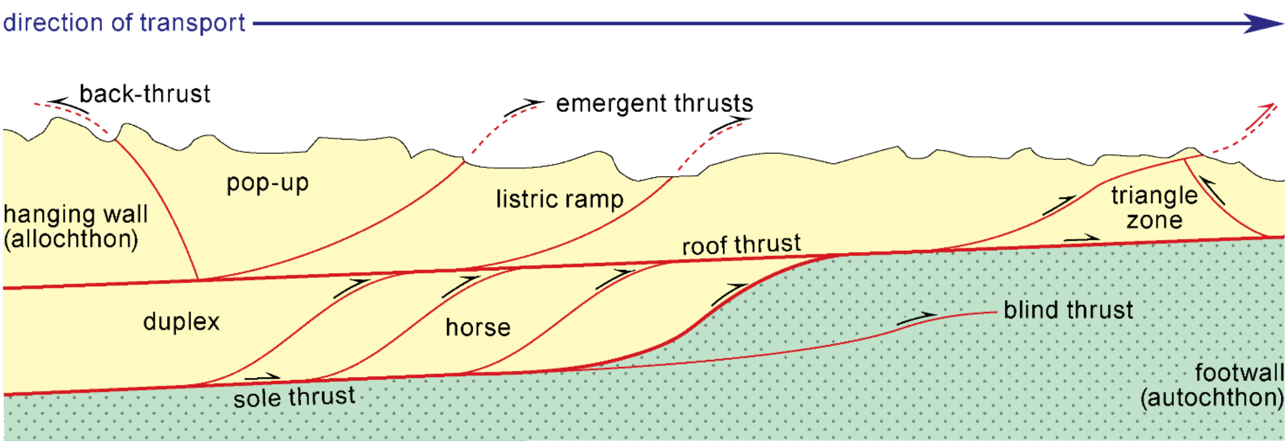


More precise definitions

Flat and ramp are defined with respect to strata orientation. Hence descriptions must indicate whether the fault attitude refers to hanging wall or footwall layering. This is done by specifying hanging wall ramp and hanging wall flat versus footwall ramp and footwall flat.

Subsidiary thrusts

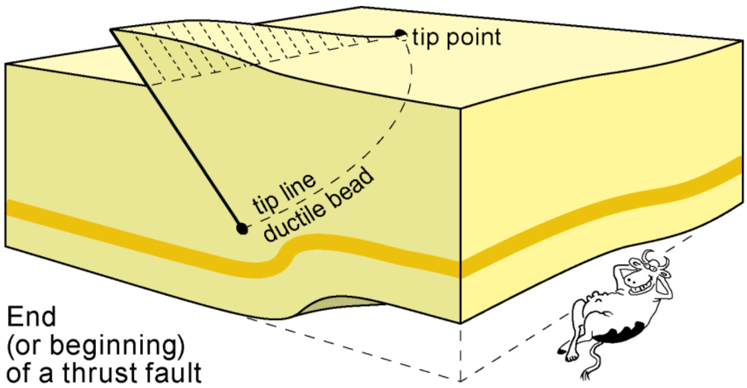
Subsidiary thrusts usually splay upward from a flat thrust. These **splay faults** (ramps) are often listric and merge asymptotically into the flat, major thrust. Subsequent tile-like piling of subsidiary thrust sheets is an **imbricate** structure.



> **Warning:** Extensional ramps cut down section in the direction of transport and are more properly termed **detachments**.

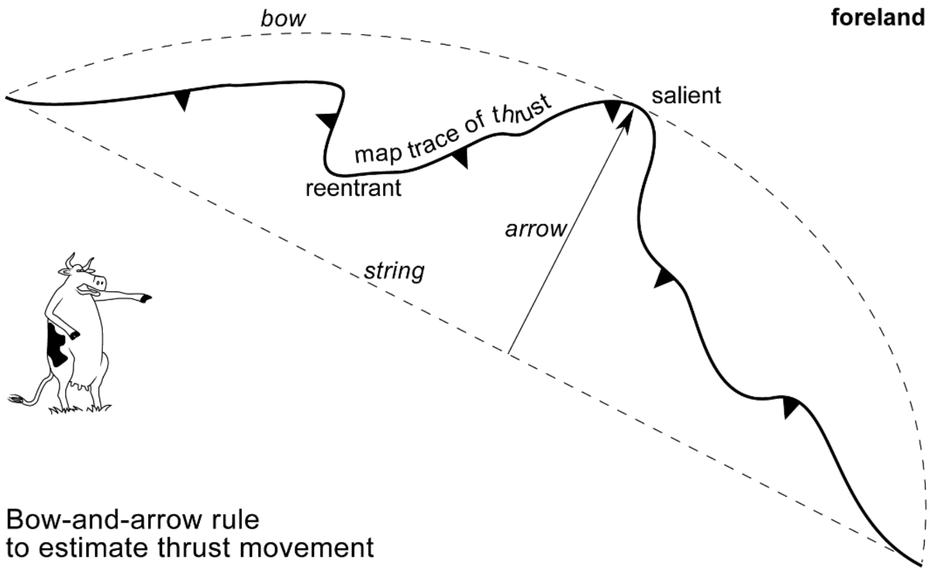
Tip line

Thrusts lose displacement in the direction of transport. Eventually, the fault displacement dies out to zero. This occurs where coherent, internal strain through the solid rock and/or folds accommodate shortening in the **ductile bead**. The termination line of the thrust (as of any other fault type) is its **tip line.** Extremities of a tip line in map view are **tip points**. In three-dimension, the termination line must be continuous and forms a closed line around the fault surface. Around the tip line and beyond the tip points, the displacement is accommodated by coherent deformation through the solid rock, the **ductile bead**.



Map trace

Because thrusts are generally shallow-dipping, their outcrop pattern may be very sinuous.
In a **salient** or **virgation**, faults and folds form an arcuate belt convex toward the foreland. The salient is ahead of the bulk main thrust front.
In a **reentrant** or **syntaxis**, the arcuate belt is concave toward the foreland. It is behind the bulk main thrust front.
Relatively high areas, or **culminations**, are usually present along virgations, and relatively low regions, or **depressions**, accompany syntaxes. However, this difference in elevations between salients and reentrants is not systematic and the contrary is known.
Large thrust faults are commonly curved in map view, typically convex towards the movement direction. This arcuate shape, imposed primarily by differential advance of the thrust-front from zero at tip points to maximum somewhere along the fault trace, is the basis for the **bow-and-arrow rule**. The movement direction is inferred to be normal to the straight, "string" line that connects the two tip points of the "bow" thrust. The thrust movement is in the direction of the imaginary "arrow". The amount of displacement varies along the "bow" and it is believed that the maximum amount of "arrow" displacement is ca 10% (± 2-3%) of the strike length. However, this assumption requires that the thrust-sheet undergoes no vertical-axis rotation of transport and no arc-parallel extension during thrusting.

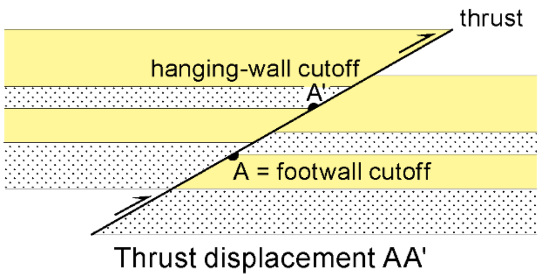


Bow-and-arrow rule
to estimate thrust movement

Tear faults (see lecture on strike-slip faults), parallel to the movement direction (i.e. nearly normal to fault strike and fold trends) take up differential displacement between adjacent segments of thrust faults. They may affect the footwall or the hanging wall only.

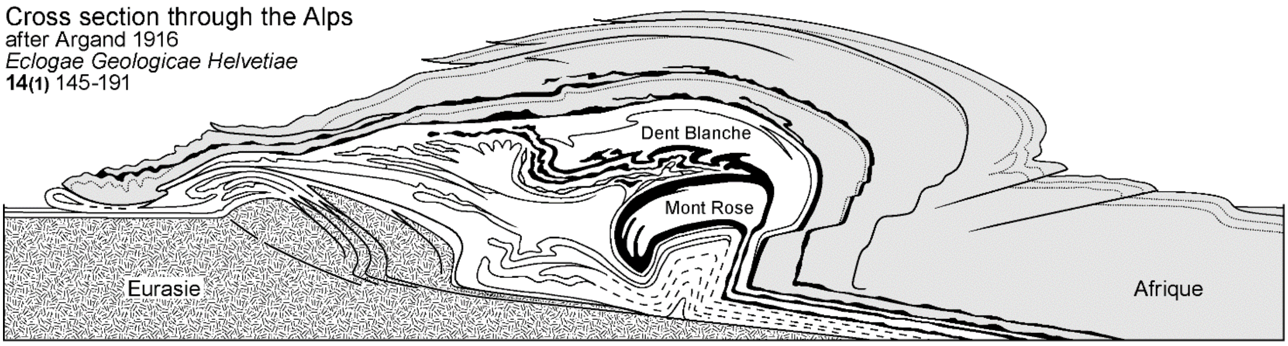
Cut-off points and lines

The intersection between a particular contact (e.g. a bedding plane) and a fault plane is a **cut-off line**, which is a **cut-off point** in cross-section. For a given stratigraphic surface, there is a **footwall cut-off** point and a **hanging-wall cut-off** point. The amount of displacement on the given section is the separation between–footwall and the corresponding hanging-wall points. The angle between the section and the slip direction must be known to calculate the actual displacement.



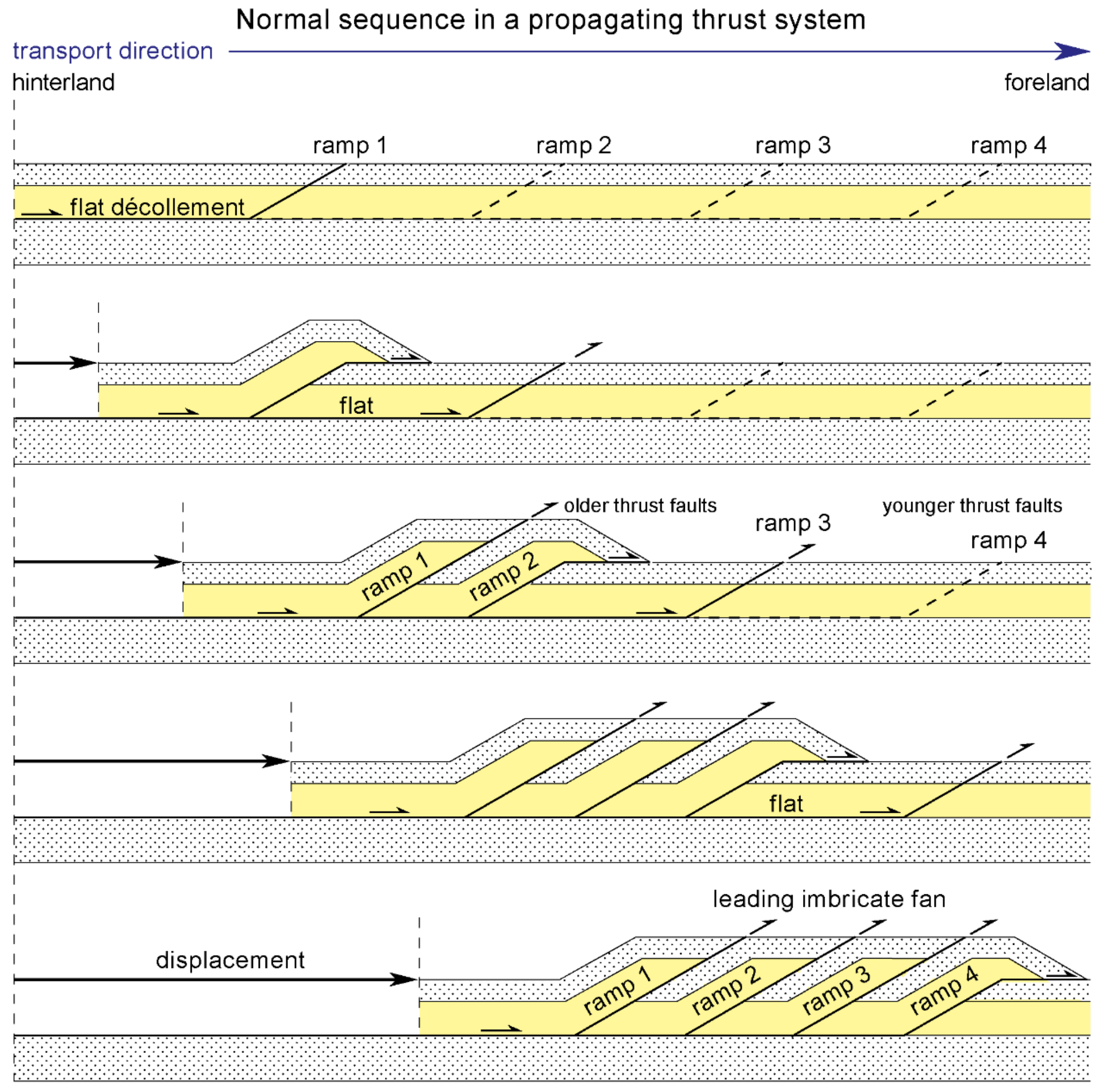
Thrust propagation

Most of our understanding on thrust propagation stems from studies in **fold-and-thrust** belts, these areas of associated folding and thrusting usually occupying a marginal position with respect to the metamorphic **axial zones** of orogens. In fold-and-thrust belts one may study the important processes that control shortening of the upper (essentially sedimentary) crust.

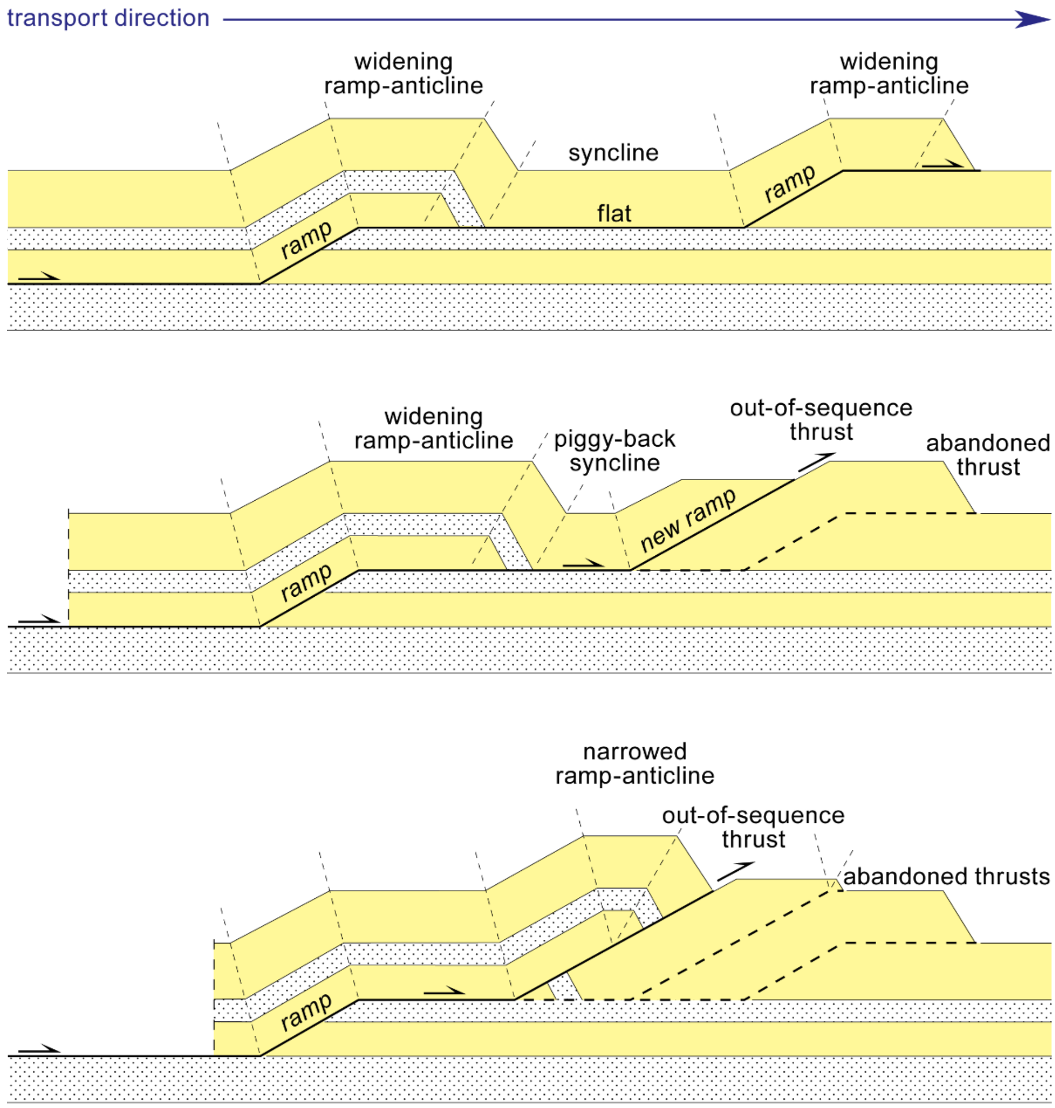


Sequence

Thrusts commonly propagate and cut up-section in the direction of slip. The major, lower décollement is initiated and climbs over a first ramp to an upper flat (which can be the topographic surface), thus bringing older (or from deeper) rocks over younger (shallower) rocks. The first ramp is deactivated when the thrust movement is transferred to the front of the system over a second ramp climbing from the same décollement surface. A third and additional ramps then form successively forward as long as shortening must be absorbed. Accordingly, younger, **normal sequence** thrusts form one after the other from the hinterland toward the foreland. In this way, the thrust system grows at the expenses of the foreland, new ramps cutting into foreland areas while older ones are abandoned.



However, new thrusts may develop back from the frontal thrust or in a random succession within the already thrust-faulted and folded hinterland. Slip on new or reactivated thrusts behind the front of a thrust system is “**out-of-sequence**”.



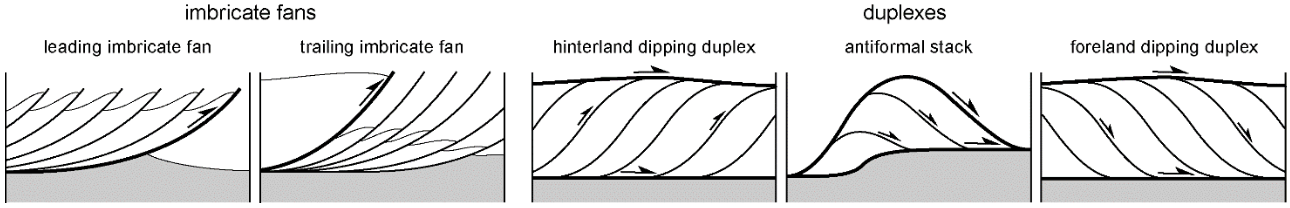
Out-of-sequence thrusting

These definitions implicitly mean that a master thrust flat is permanently active during the shortening history, while most of associated thrust faults are transient structures with a short lifetime.

Imbricate fans

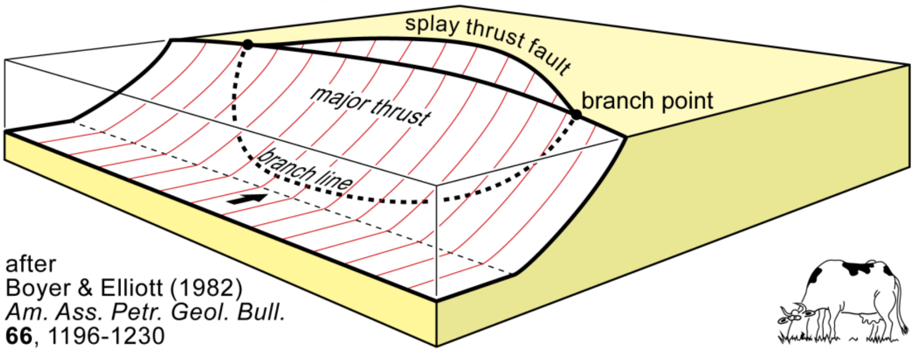
Faults with large displacement commonly die out in a set of smaller, commonly sub-parallel **splay faults**. They normally form sequentially as the location of the frontal fault surface jumps ahead to ramps that cut into the foreland.
Splay faults branching off and ramping all in the same direction out of the main, deeper décollement make an **imbricate fan**; the fan spreads the major displacement over a large volume of rock.

Different systems of thrusts
after Boyer & Elliott (1982) *Am.Ass.Petr.Geol.Bull.* **66(9)**, 1196-1230



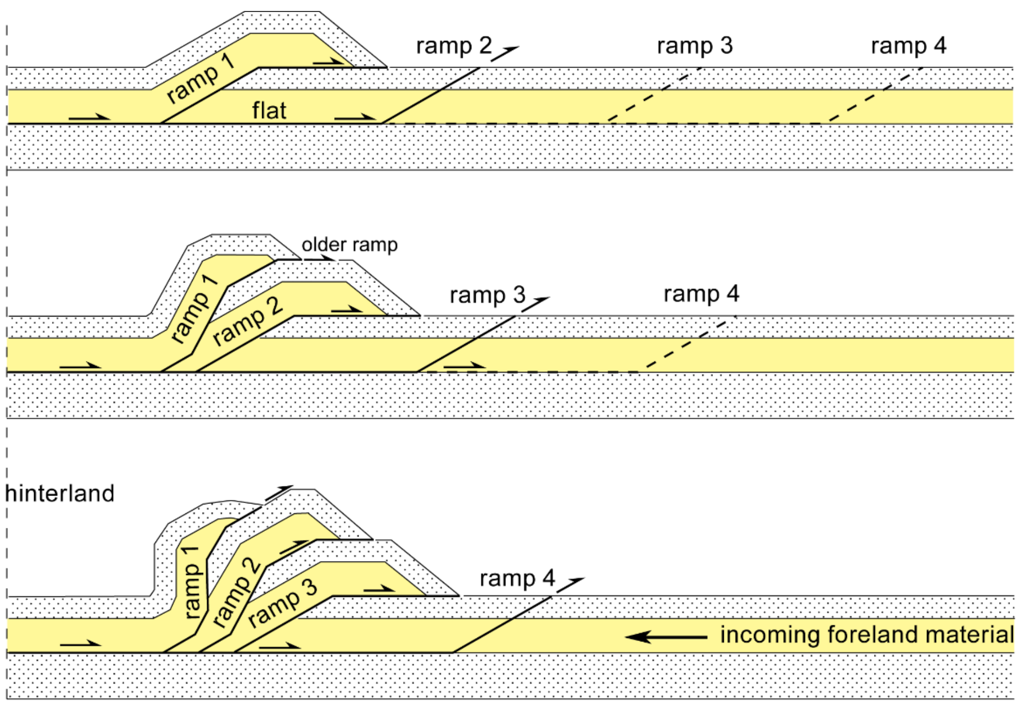
The junction line where the main thrust fault splits into two smaller thrust surfaces is a **branch line**. The **branch points** are where the traces of two thrusts meet on a cross-section.

"Horse" terminating at a branch line and two branch points



Attention: do not confuse with the intersection line of two non-contemporaneous fault planes, the older against the younger.

The individual thrust sheets of the imbricate fan are **schuppen** (fish-scales in German). The branch points at the floor of an imbricate fan, where two thrusts separate toward the foreland, are called **trailing branch points**. The merging point of two thrusts joining into one as they are traced toward the foreland are called **leading branch points.**
The youngest splay is the front thrust in a **leading fan.** Maximum displacement is absorbed in the frontal thrust. In that case, the youngest splay carries the older one "on its back", which is called **piggy-backing**. Stacking of new imbricates at the frontal base of the fan progressively steepen the older splays and horses through passive rotation to the back.



Steepening of older horses due to stacking of younger imbricates at the imbrication front

The youngest splay is at the rear in a **trailing fan**. In that case there is thrusting over older splays. One of these reactivated splays absorbs maximum displacement.

Piggy-back transport

Where the later thrust develops in the footwall of the original thrust, structurally higher and older thrusts and hanging-walls are carried forward in a **piggyback** manner by lower, younger thrusts. Conversely, if the thrusts migrate backwards, an **overstep** sequence develops.

Tectonic triangles and wedges

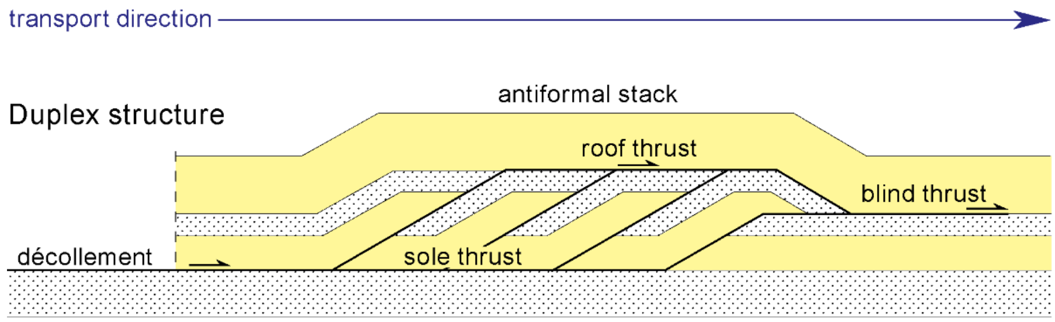
A block of rocks contained between a pair of oppositely moving, conjugate thrusts is a **tectonic wedge**. It is a **pop up** when the conjugate thrusts and back thrusts diverge upward, thus tend to move the block upward. The tectonic wedge is a **triangle zone** when the bounding thrusts and back thrusts diverge downward, thus tend to push down the block**.**

A **tectonic wedge** is the leading edge of a thrust unit between a pair of oppositely moving but merging fault planes (imagine entering a crocodile mouth).

Duplex structures

Definition

Thrust systems commonly involve several approximately parallel décollement-surfaces whose location and extent are controlled by weak layers at different levels of a sedimentary pile. A **thrust-duplex** consists of a series of sub-parallel ramps that branch off a relatively flat, lower **floor thrust** (also called **sole thrust**) and merge upward into the upper **roof thrust**. The whole structure encloses a package of S-shaped, detached slices of rock stacked in a systematic manner. The individual imbricate lenses are called ***imbricates*** or **horses**. Typically, horses make a progressively larger angles with the roof- and floor-faults from front to back (like in leading fans). Unlike an imbricate fan, a thrust duplex is contained within the sedimentary sequence.



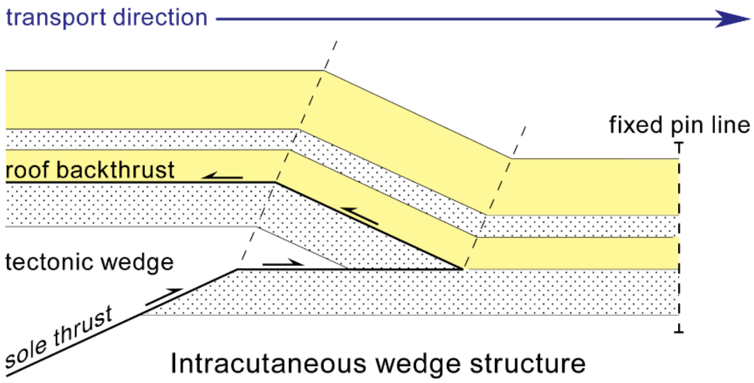
Development

Duplex formation is initiated when the forward propagation of a thrust is impeded by some perturbation or sticking point. The thrust is forced to ramp up to a higher glide horizon. With continued displacement on the thrust, higher stresses are developed in the footwall of the ramp, which makes an obstacle to the horizontal movement of rocks. Increased stresses cause renewed propagation of the floor thrust ahead of the ramp along the décollement horizon, until the fault plane again cuts up to join the roof thrust. Further displacement then takes place along the newly created ramp. This process may repeat many times, forming a series of fault bounded, typically a lozenge shaped horses. The tectonic shift of the footwall ramp by the sequential formation of thrust slices creates the duplex structure. The development of each new thrust slice is accompanied by the backward rotation and 'piggy-back' transport of the earlier-formed horses. Parameters that determine the final geometry of the duplex include the ramp angle, the initial and final spacing of the thrusts, and the amount of displacement on them.

Types of duplex structures

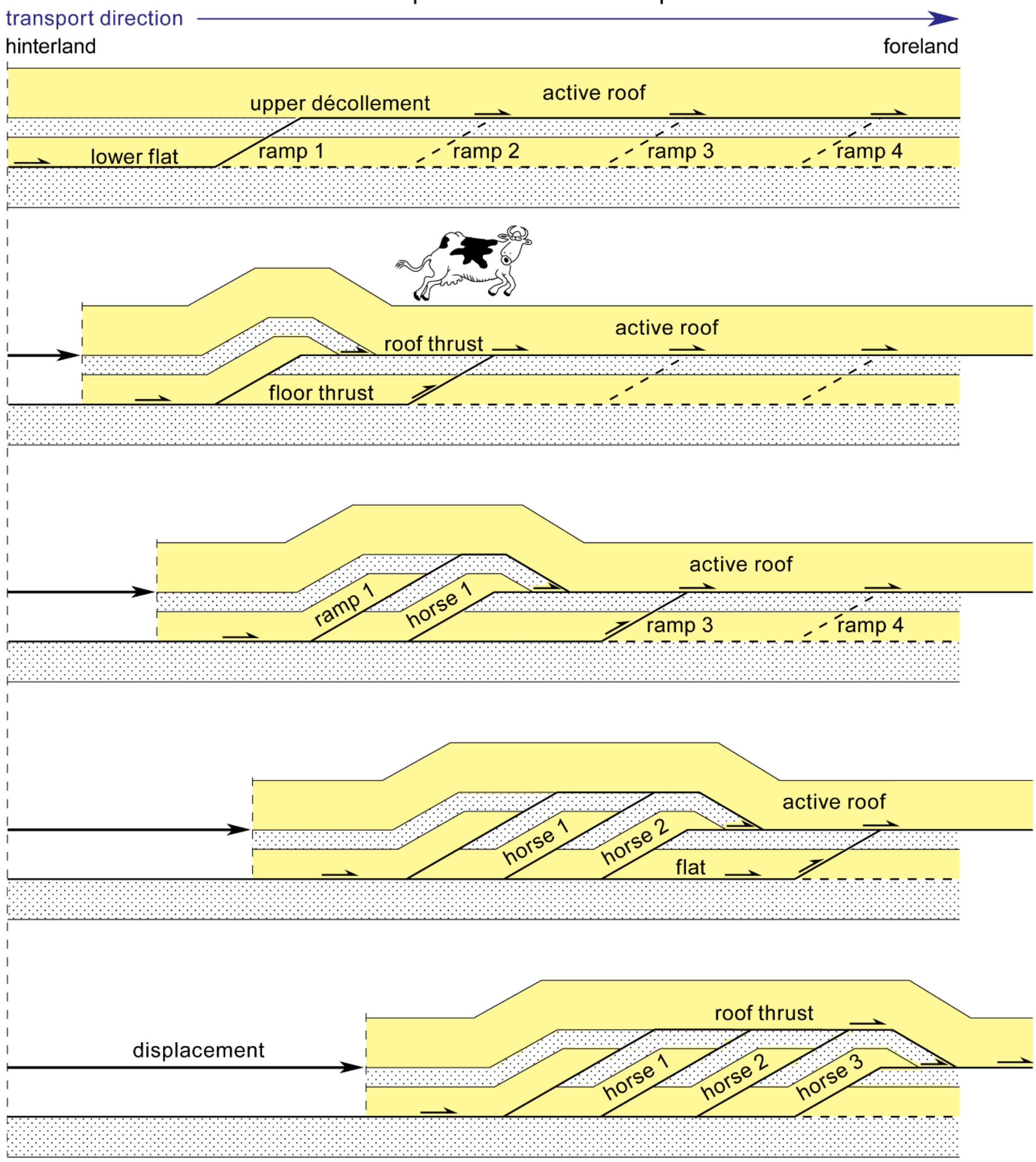
The displacement between the rocks lying above the roof thrust (the roof sequence) and the horses within the duplex defines two types of duplexes:

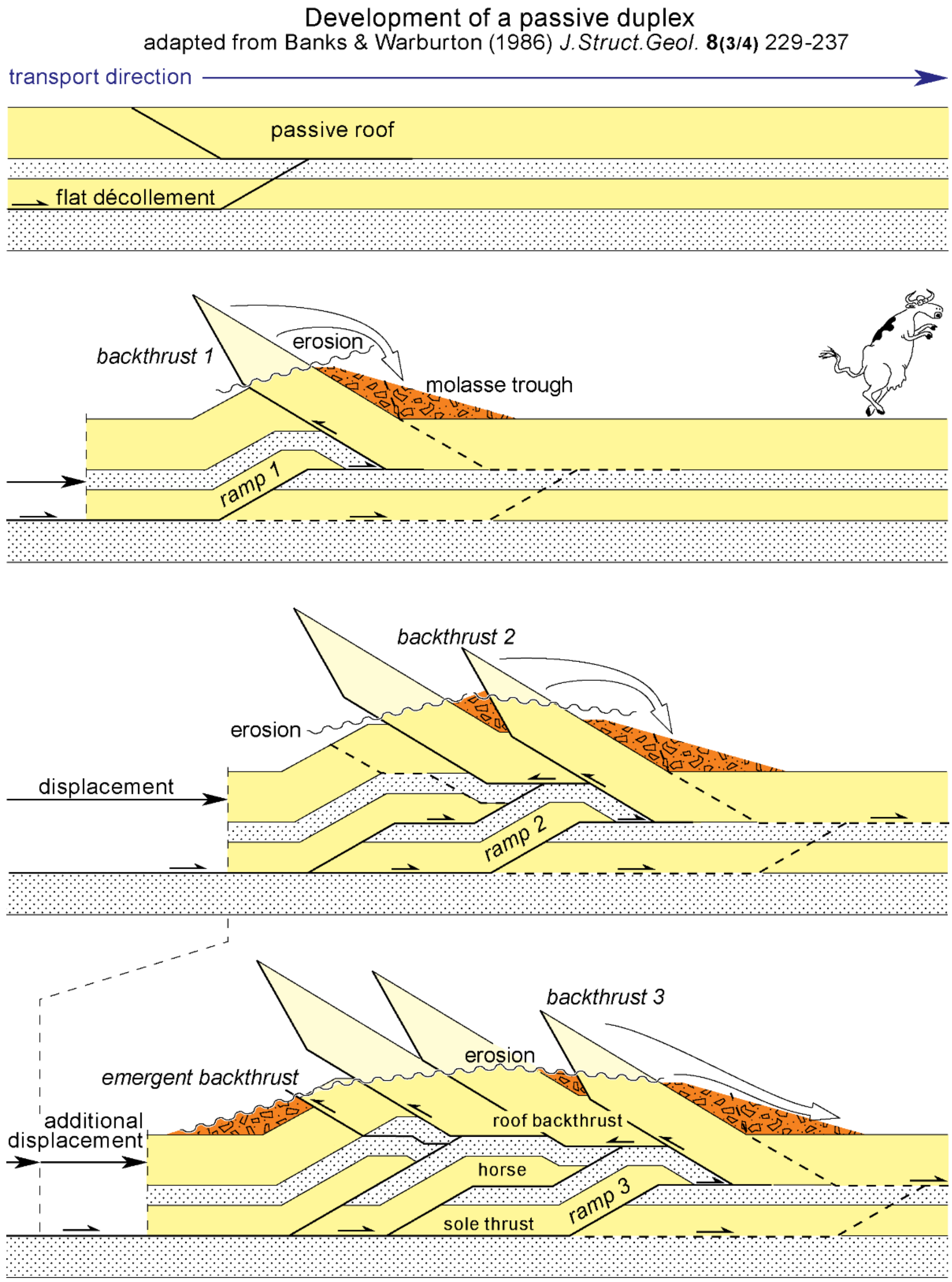
- Active-roof duplexes, where the roof sequence moves forward, as the horses.
- Passive-roof duplexes where the roof sequence moves opposite to the horses. Backthrusting produces a frontal **intracutaneous wedge structure** terminating at a buried tip line and underthrusting of the roof sequence by the horse, wedge blocks.



Intracutaneous wedge structure

Development of an active duplex

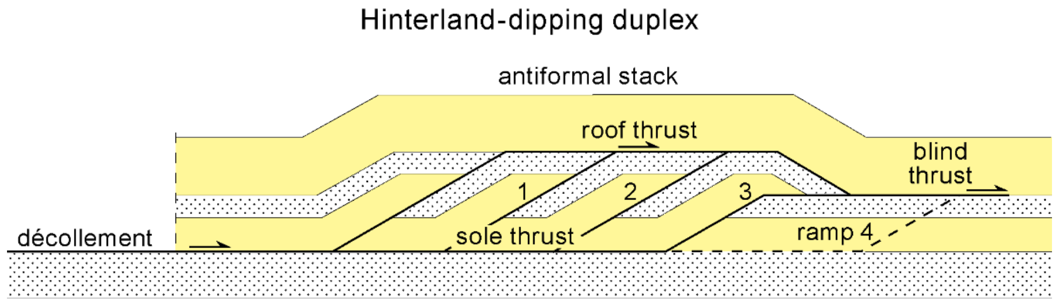




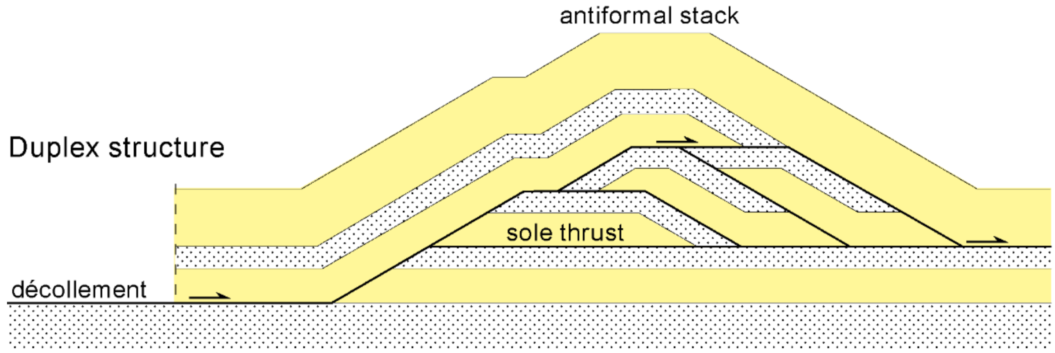
Morphology

As with imbricate fans, there are different duplex morphologies, which depend on when the new horses are formed relative to older horses and on the amount of displacement of rear horses with respect to frontal ones.
a) In most duplexes the ramps bounding the horses have relatively small displacements; new horses are formed at the front (in the slip direction);

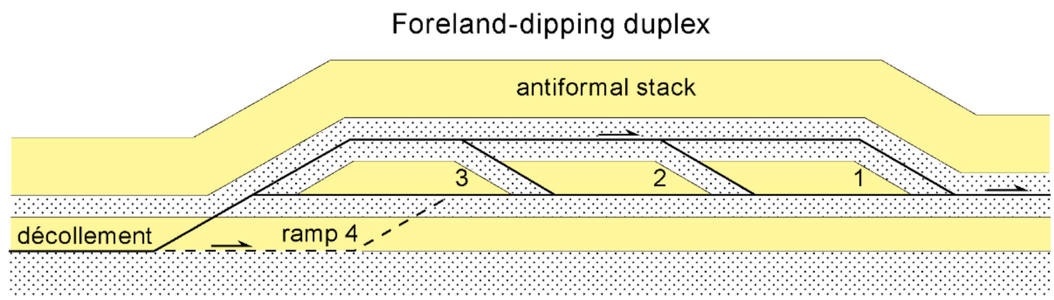
The ramps and the horses dip away from the foreland. The final geometry is a **hinterland-dipping duplex**. This is the most common type of duplexes.



b) The displacement on the individual ramps can be greater, such that horses are piled on top of each other and form an **antiformal stack**, which appears as **eyelid window** when the hanging wall is eroded.



c) If the ramp displacement is still greater, higher, older splays may be reactivated to move an older horse over and beyond the antiformal stack of younger horses. The geometry is a **foreland-dipping duplex**.



Relationship between folds and thrusts

Asymmetric, open to close, eventually overturned folds are commonly explained with thrust-propagation models, which employ migrating, kink-shaped hinges and relate fold geometry entirely to fault geometry, slip and to the thickness of the transported layers. Such models assume that faults propagate gradually and that accumulating slip, which must be zero at the front tip, is largely absorbed in contemporaneous folding. However, other models show that folding precedes thrusting and that the periodicity of thrusts is inherited from the buckling wavelength of the earlier folds.

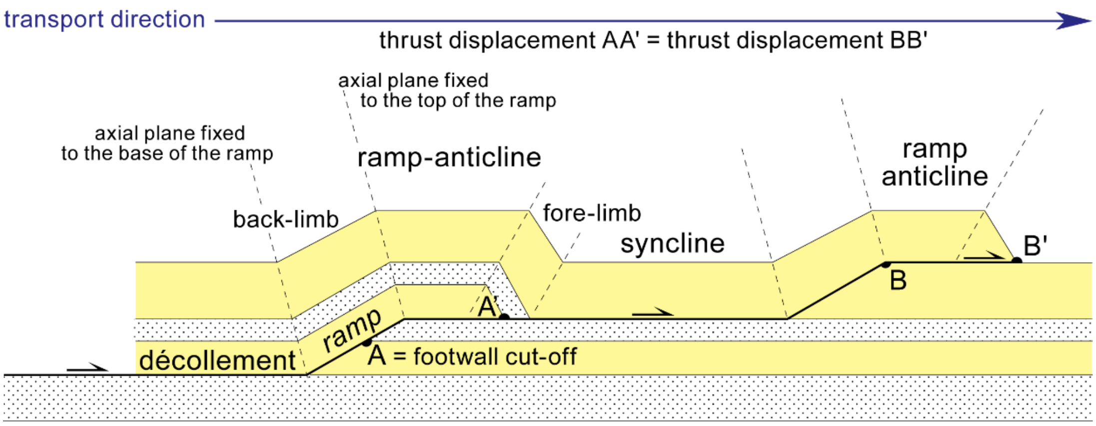
Folded thrusts

Because folding and thrusting are closely linked in most shortening processes, it is quite common for an originally low-angle thrust fault to be rotated either:
- into a steep orientation (reverse fault).
- into a flat-lying orientation where the hanging-wall has actually moved down (geometrically a low-angle normal fault).

In either case, to avoid ambiguity, the thrust is a fault that puts older rocks on top of younger. Notice also that for the most part, thrusts cut bedding at lower angles than normal faults.

Passive folds

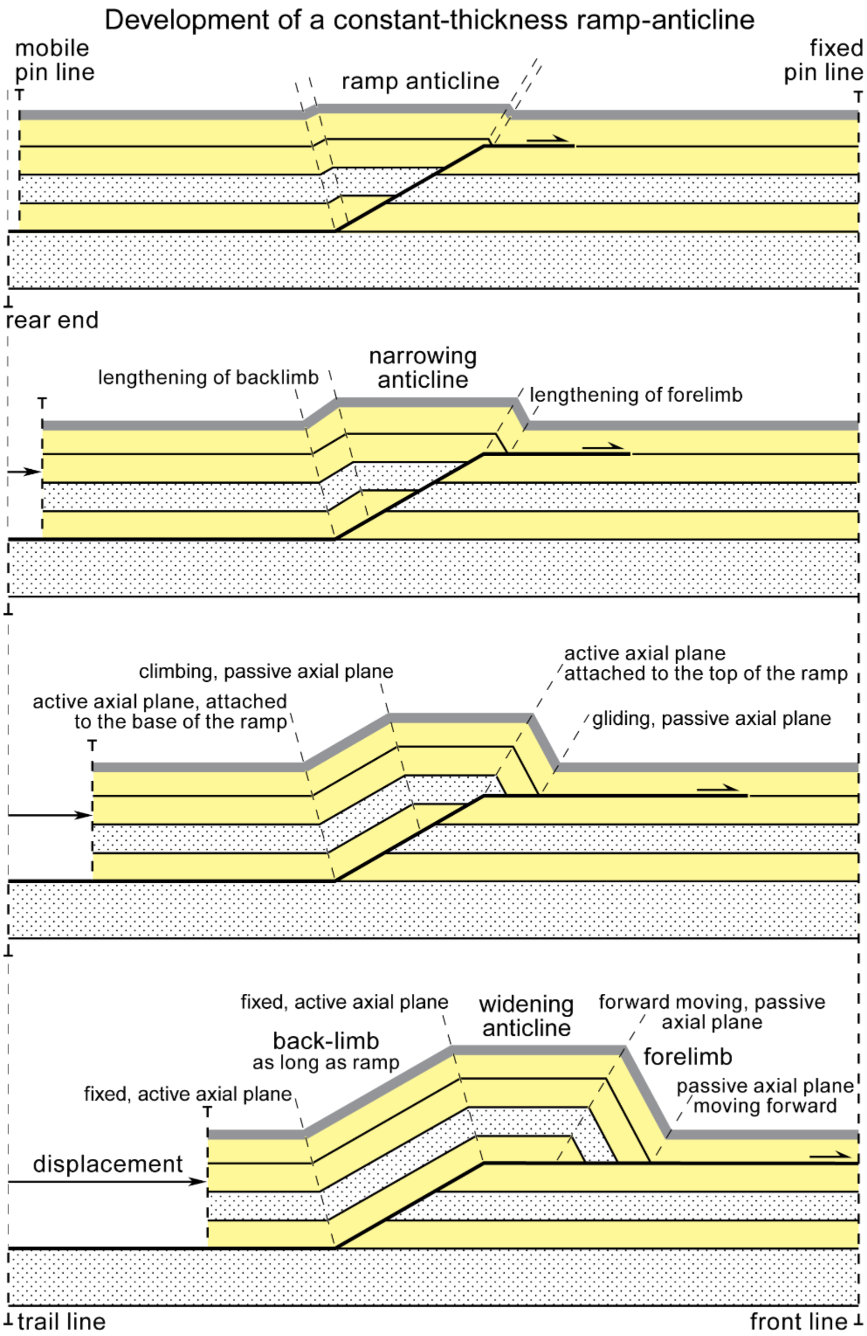
Folds grow at the tip of a blind thrust fault where propagation along the décollement has ceased while displacement on the thrust behind the fault tip continued. Typically, this continued displacement is accommodated by an asymmetric anticline-syncline fold pair. Folds also form when layers move from flat to ramp or vice versa and must conform to the fault geometry.



Geometrical rules

Most passive folds in thrust belts are parallel folds, meaning that bedding thickness is fundamentally preserved. Most of these folds consist of sharp and narrow hinge zones separating flat limb **panels** of approximately equal dip. Cross-section construction with a kink-like fold geometry generally yields very satisfactory solutions. The geometrical rules often applied to infer the thrust geometry (unexposed flats and ramps) from kink band models (**kink construction**) are:

- Axial surfaces bisect the angle between the fold limbs.
- Axial surfaces terminate downward at the bends (flat ↔ ramp transitions) of the thrust plane.
- Where two axial surfaces intersect, a new axial surface is formed, also satisfying the equal-angle rule.
- In the kinematic model of fault-related passive folds, hinges must be mobile during fold growth. There are two types of axial planes:
    - (1) Active axial planes are fixed relative to the fault ramps and flats. Each bend in the fault is associated with an active axial plane through which material moves.
    - (2) Passive axial planes are fixed relative to the layers they bend. They move together with the material along the fault.
- The thickness of each layer remains constant throughout the structure, except in the forelimb of the fold to allow the leading-edge triangle of the hanging wall to rest against the underlying thrust plane.
- Thickening or thinning must be constant throughout the forelimb; there is a strict relationship between the dip of the ramp α, the interlimb angle 2δ and the thickness change occurring in the forelimb t_0/t_f depending on the fault type.
- The maximum amplitude is the height of the step in ramp.



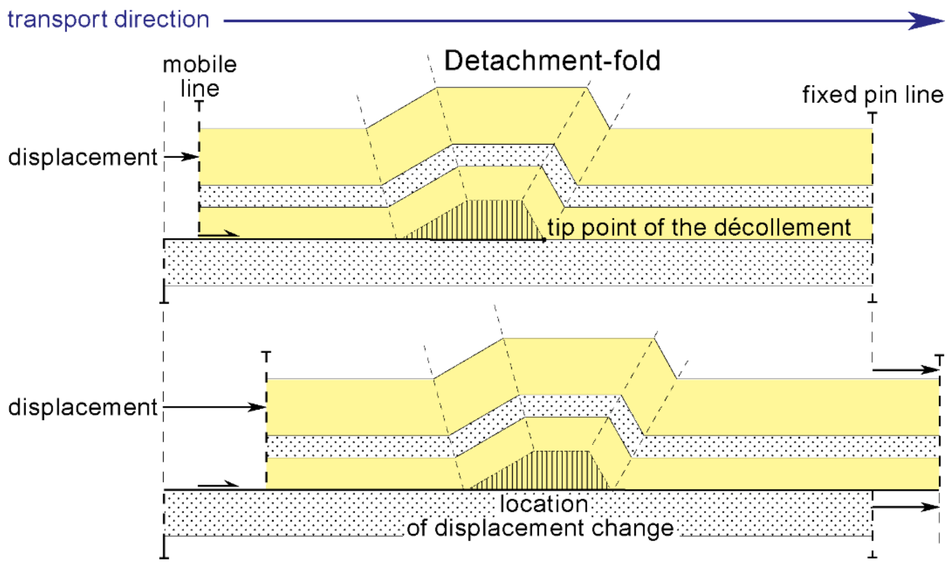
Three geometric types of thrust-related folds are recognized: **detachment folds**, **fault-propagation** (or **tip-line**) **folds** and **fault-ramp** (or **fault-bend**) **folds**. Note that they represent three different (eventually successive) stages of the development of a ramp.

Detachment folds

Description

Folds occur where the amount of horizontal displacement of the hanging wall on a flat, blind décollement changes. For example, folding-deformation is geometrically required to compensate the absence of fault movement in front of the tip of the fault plane. Like a carpet buckles if one pushes one side on the floor, relatively competent layers buckle above a bedding-parallel décollement, which is contained within an incompetent (weak), often disharmonically folded layer.
Amplification of the lift-off anticlines expresses the upward escape of the incompetent, ductile décollement material heterogeneously thickened in the fold core to accommodate the displacement gradient. This displacement diminishes progressively along the flat in the foreland direction,

eventually down to zero at the flat tip, which anchors the frontal axial plane. Depending on the behaviour and thickness of the incompetent décollement layers, detachment folds have various shapes from asymmetric kink-like anticlines to nearly symmetric box folds. The fold becomes smaller downward, towards the tip of the décollement (axial planes converge).

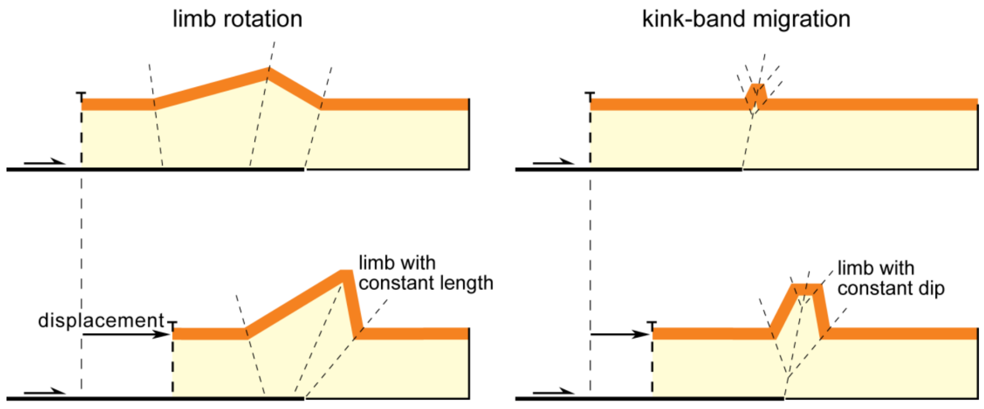


Kinematics

Two end-member mechanisms control the formation of detachment folds:

(1) Rotation of the limbs, making the fold ever taller and narrower. The length of the limbs remains constant and the anticlinal hinge (kink) remains on the same material point within the folding competent (lid) layers.

(2) Migration of the kink band. The fold limbs maintain a constant dip angle but become longer with progressive fault movement.



End member mechanisms for the development of detachment-folds

In reality, these two end-member mechanisms are often combined.

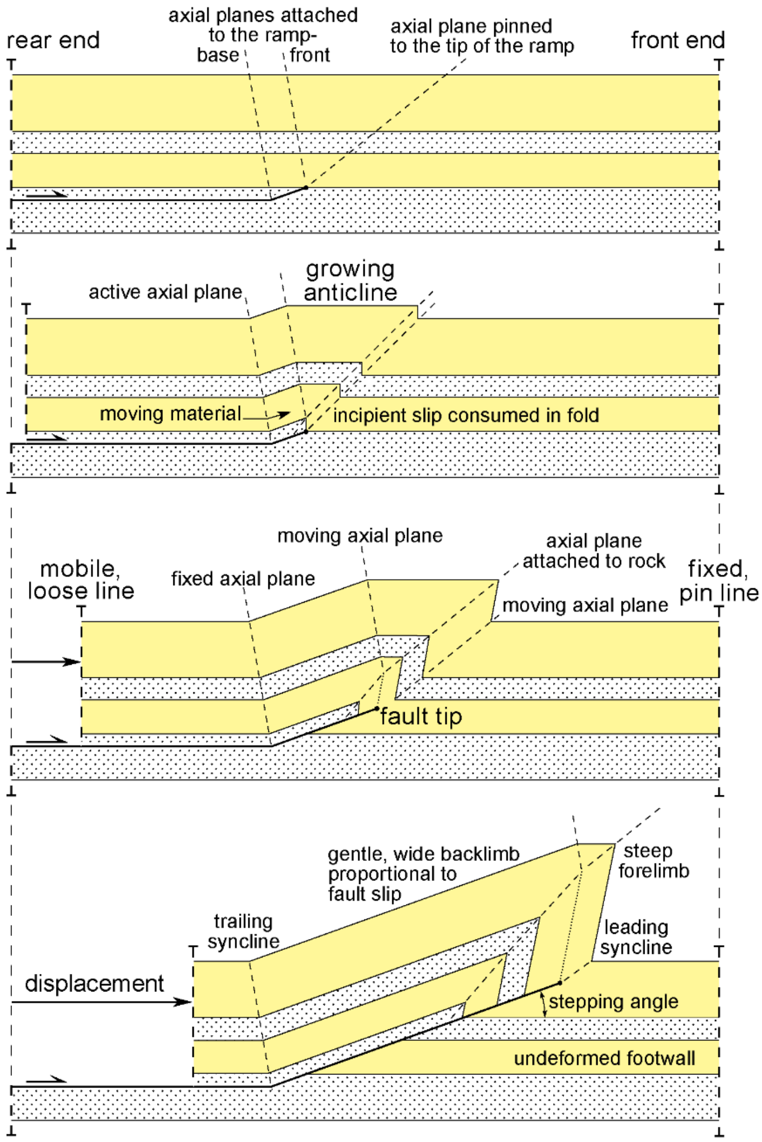
In all cases, the frontal axial plane separating the forelimb from the flat, undisplaced foreland is pinned to the tip of the blind décollement. The forelimb is commonly steeper than the backlimb; the fold asymmetry is consistent with the transport direction. This asymmetry is a direct consequence of the diminishing -slip towards the foreland: there is less slip/thickening to absorb on the front-side of the fold than on the back-side.

The folds of the Jura Mountains, where the floor décollement propagated within Triassic evaporites, are a classical example.

Fault-propagation (tip-line) folds

Description

Blind thrusts often terminate upward in splays from a flat décollement, forcing layers to bend ahead of the propagating fault tip as material moves up the ramp. Folding is coeval with thrust propagation. The results are markedly asymmetric **fault-propagation folds** whose shape depends on the amount of displacement along the basal décollement, the ramp inclination and the slip to propagation ratio. Fault-propagation folds are tighter downward.



Kinematic evolution of a fault-propagation fold

Kinematics

An anticline grows to consume the increasing amount of slip on the forward and upward propagating ramp. Layers bend forward around the fault tip and the total displacement is absorbed by the frontal limb until the fault breaks through all layers. The limbs are bounded by axial planes, i.e. kink band boundaries in kink-geometry.

- The rear axial plane is pinned to the base of the ramp; it bisects the flat-ramp angle and layers form a synform as they pass through this "active" plane to climb the ramp. This ramp-bottom axial trace remains there and is active as long as material passes by.

- The frontal axial plane is pinned to the propagating tip of the thrust plane; it is bisecting a syncline whose limbs are the undisturbed layer on the front side and the steep, occasionally inverted limb tilted forward to absorb the slip along the thrust fault. The asymmetry of the fold is obviously consistent with the direction of local thrust displacement.

In between, the fault-propagation anticline is halved by an axial plane abutting downward against the ramp. The upper tip of this axial trace lies on the same bedding plane as the fault tip. The anticlinal hinge is a point (in kink geometry) from which two axial traces diverge upward.

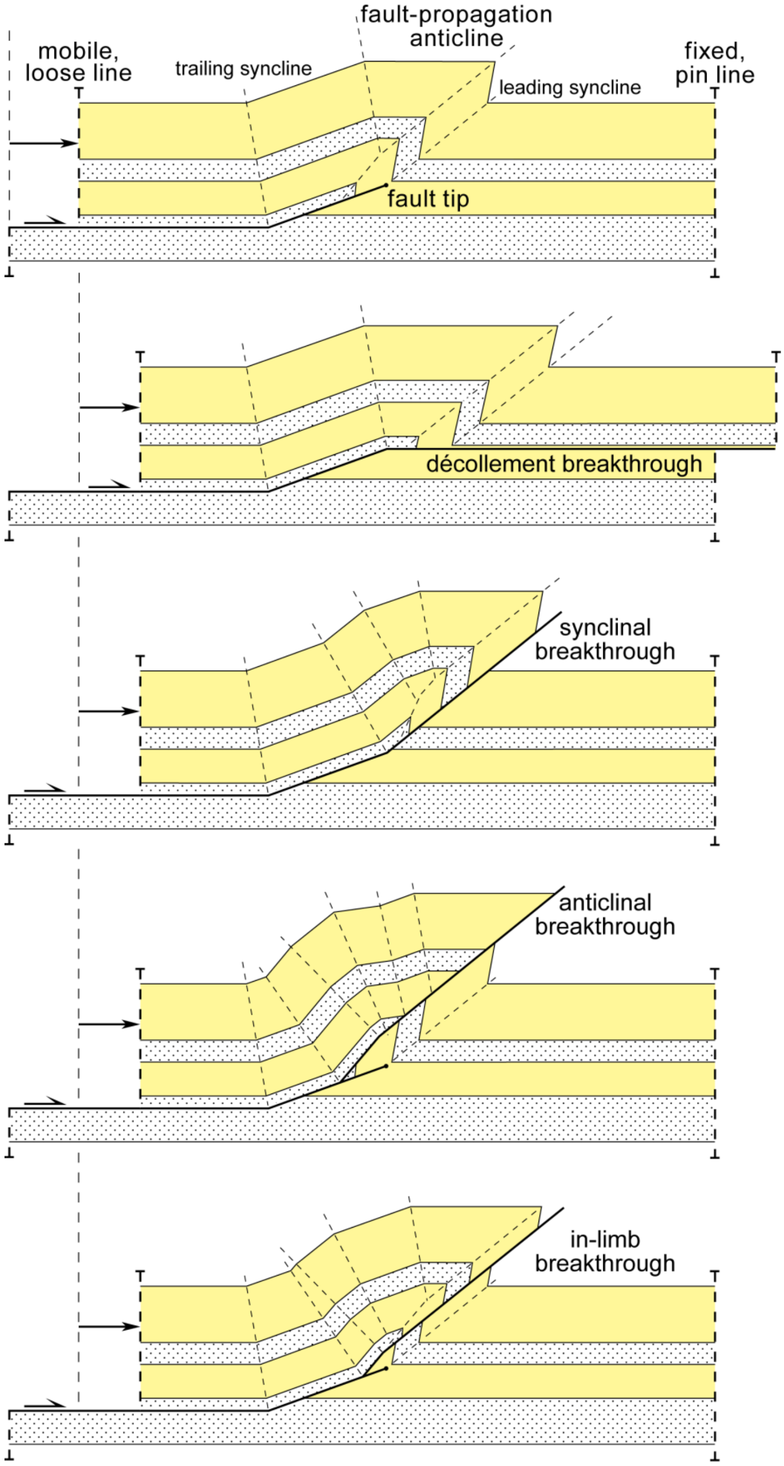
- One is parallel to the rear one; these two twin axial planes define the rear limb, parallel to the ramp.
- The other is parallel to the frontal axial plane; these two axial planes define the front limb (kink band) of the fault-propagating fold.

The anticlinal axial plane and the ramp lengthen during fault propagation folding (kinking). The splitting point of the axial planes of the anticline must move upward and forward to remain on the same stratigraphic plane as the tip of the ramp. The rear axial plane stays anchored at the bottom of the ramp. All other axial surfaces are active and move through the material. The two divergent planes, maintain their original orientation and remain attached to the tip of the axial plane of the growing anticline, which also moves up the ramp. Consequently, limbs lengthen while the fault tip is propagating forward.

Exercise

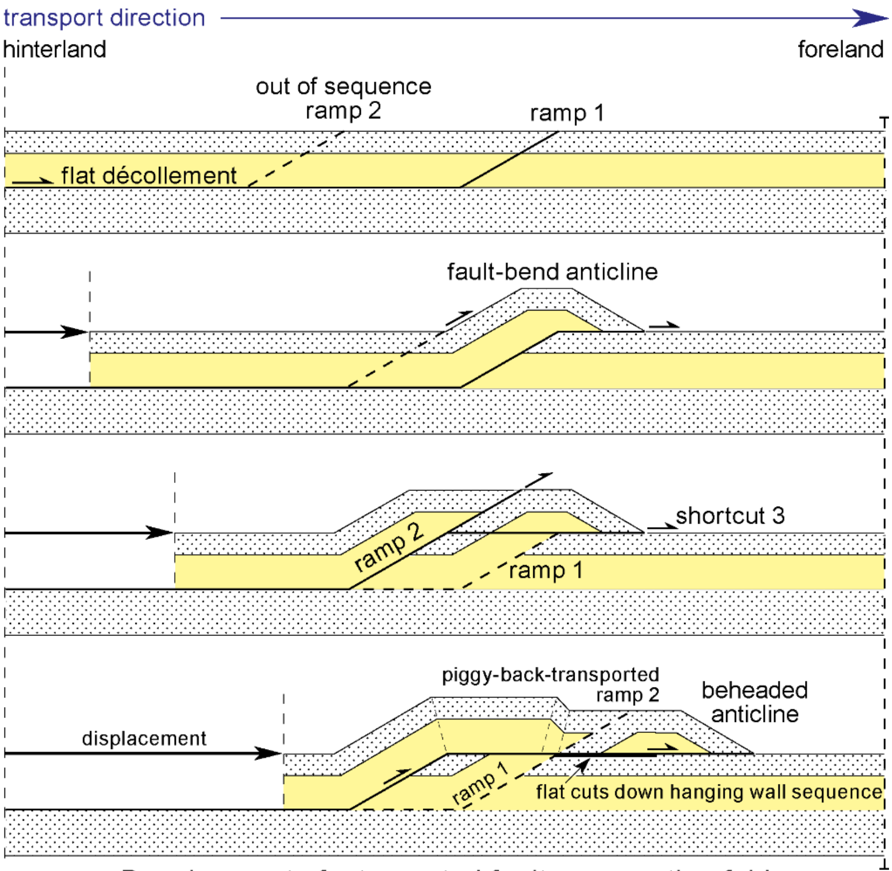
Using kink geometry, draw and study the development of a fault propagation fold.

Commonly, fault-propagation folds become locked because the bending resistance of layers is too big. The thrust plane may break through along the anticlinal or synclinal axial surfaces or somewhere in between within the steep limb to further follow-a flat décollement following a weak datum.



Types of breakthrough structures

The thrust plane may propagate beyond the area of folding and eventually truncates and shears off fault-propagation folds whose development is choked. In this case the propagating fault leaves truncated folds in the hanging-wall.



Development of a truncated fault-propagation fold

Fault-ramp (-bend) folds

Fault-ramp folds develop where a blind thrust ramps up from one flat to a higher level flat. Folding postdates the thrust development.

<u>Description</u>

Material displacement over non-planar thrust planes induces distortion of the hanging-wall. The geometric characteristics of this distortion depend on the orientation and size of the fault topography with respect to the transport direction. The trend of the resulting fault-ramp folds reflects the strike of the ramp below the thrust sheet. The forelimb of fault-ramp folds is always located on the foreland side of their associated ramps. The rear limb is parallel to and behind the ramps.

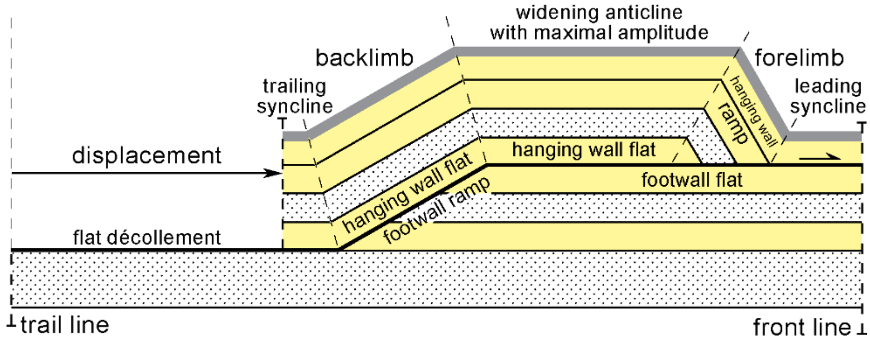
<u>Kinematics</u>

Whereas fault-propagation folds develop simultaneously with, and immediately above the propagating ramp, fault-ramp folds develop subsequent to the ramp formation. Hanging-wall rocks are tilted parallel to the inclination of the ramp as they ride up the ramp. They recover their initial attitude once they passed the ramp.

- The rear axial plane is pinned to the base of the ramp; it bisects the flat-ramp angle and layers form a syncline as they pass through this “active” plane to climb the ramp.
- With thrusting appears a new axial plane parallel to the early rear one from which it instantaneously separated with the material points initially on the active axial plane and passively climbing the ramp; these two parallel “twin” axial planes delimit the rear limb, parallel to the ramp. The length of this limb, i.e. the distance between the twin axial planes, is proportional to the amount of thrust slip.
- A third axial plane emerges immediately from the top of the ramp, where material moving from the ramp onto the upper flat must bend down and forward to conform the upper flat. This axial trace dips towards the hinterland, with an angle that allows respecting a constant bed-thickness / layer length geometry in parallel folds (and kinks).

- A frontal axial plane, parallel to and coeval with the third one, is pinned to the frontal tip of the thrust plane. This frontal plane bisects a syncline whose limbs are the undisturbed layer on the front side, and the limb tilted forward in front of the third axial plane. For geometrical reasons, the fore limb generally dips at a higher angle than the back limb. The asymmetry of the **ramp-anticline** (also called rootless anticlines) is obviously consistent with the direction of local thrust displacement.

Passive transportation of hanging wall layers in fault-ramp folds



The development of fault-bend folds is somewhat comparable to the history of fault-propagation. A syncline-anticline pair is formed: the syncline above the basal transition from flat to ramp, and the anticline above the top transition from the ramp to the upper flat. The rear axial plane is fixed to the bottom of the ramp and bisects the flat to ramp angle. The other axial planes are more mobile.
- The rear "twin" axial plane climbs the ramp while remaining parallel to itself, until it reaches the top of the ramp. At this point, the ramp anticline reaches its maximum amplitude and the climbing axial plane stops to remain anchored at the ramp top.
- The moving material permanently entrains forward the frontal axial trace.
- When the rear "twin" axial plane reaches the ramp top, the axial plane previously attached to the ramp-top becomes passive and slips forward on the upper flat with the frontal axial plane. Then the ramp anticline widens without heightening of structural relief as long as there is thrust displacement. The final anticline amplitude is the thickness of the hangingwall sequence.

Exercise

Using kink geometry, draw and study the development of a fault-bend fold on a ramp between a lower décollement and a frontal, upper flat.

If the displacement has a component down the ramp, then a syncline develops.

Imbricate structures

In some imbricate thrust systems, horses are bunched up in the form of an **antiformal stack**.

Exercise

Using kink geometry, draw a hinterland dipping duplex (small displacement) an antiformal stack (intermediate displacement) and a foreland dipping duplex (large displacement).

Practical tip

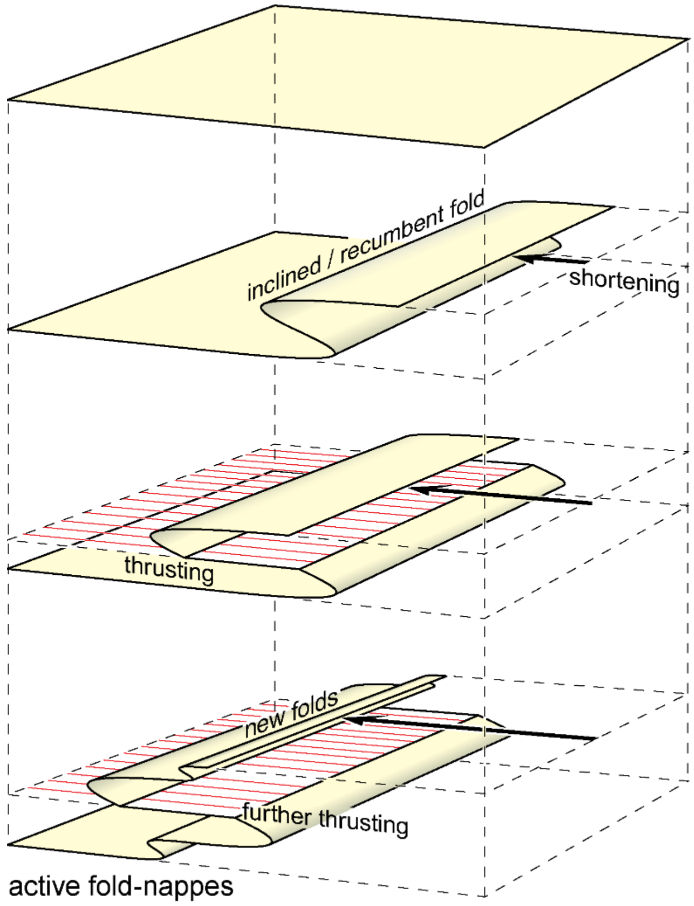
Layers of the hanging wall are almost everywhere parallel to the footwall layers. Yet, sketches of the exercises show that:

- Hanging wall layers cut footwall layers along ramps.
- Footwall layers cut hanging wall layers at the base of forelimbs of ramp-related folds

These relationships help knowing where in a thin-skin thrust system field observations are made.

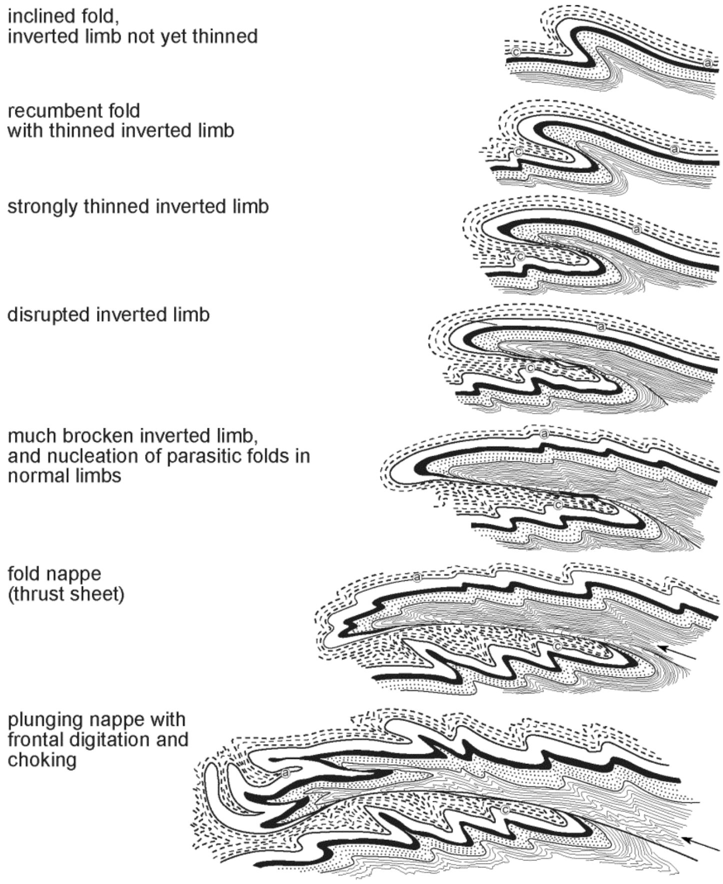
Active folds

Active folding means either that folding and thrusting processes are coeval or that folding absorbs some shortening before thrust activation. These two possibilities largely depend on the metamorphic conditions. At low temperature and pressure conditions (i.e. within the sub-surface crust) thrust-related folds are passive. Shortening strain becomes progressively more ductile with depth, i.e. folding becomes more important.



Stretched-fold thrusts (fold nappes)

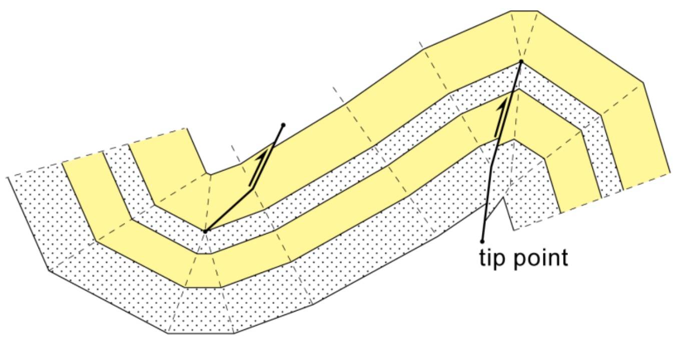
The concept has been developed in the Alps by the beginning of the 20th century. The ductile, overturned limb of a growing fold stretches and thins down until it breaks into a thrust.



Evolution from recumbent fold to fold nappe and thrust sheet
sketched after Heim A. 1919 *Geologie der Schweiz* Tauchnitz, Leipzig, 704 p.

Fold-accommodation thrusts

Subsidiary thrust and reverse faults may form in tight fold cores where bending and/or flattening become insufficient to accommodate excessive shortening. Increasing bed curvature pinches out the fold cores where strain and volume problems develop local stresses that reach the yield stress of the rocks. In that case, folding is the causal process for faulting.

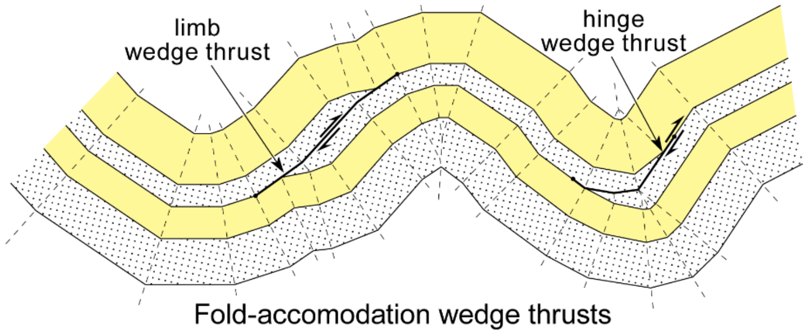


Fold-accomodation reverse faults in a syncline-anticline pair

Fold-accommodation thrusts and reverse faults have some characteristics:
(1) Largest movement is rather small, decreases rapidly and always cuts up stratigraphic section towards the core of either antiforms or synforms.
(2) Fault planes are isolated and smaller than the associated fold; they occur at different stratigraphic levels, mostly across competent layers, and terminate without linking by flats.
(3) Thrust tips form an angle with bedding and do not necessarily run into bedding-planes. Along strike, transitions between folds and thrusts are frequent.

(4) Fold axes can be followed from footwall into the hanging wall and thrust-planes are commonly deformed more or less harmoniously with associated folds.
(5) A geometric and kinematic relationship to surrounding, often strongly disharmonic folds. In particular their strike is generally parallel to the fold axes and they show a more or less symmetric arrangement with fore- and back-thrusts on either side of the fold axial planes. Slip on these conjugate faults logically accommodates excessive, bulk shortening strain.

Flexural slip involves bedding-parallel slip towards hinges. Movement planes can cut through a folding layer in a flat-ramp geometry producing wedge thrusts in fold hinges and/or fold limbs.



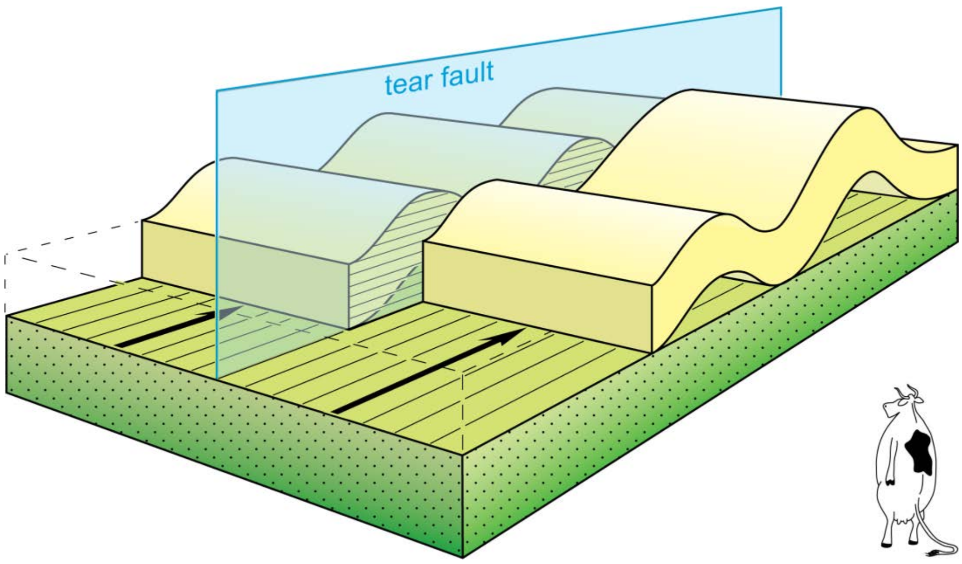
Fold-accomodation wedge thrusts

Folds-over-ramps

Folds with complex geometry may develop in hanging wall going over a ramp, in response to local mechanical instabilities.

Tear faults and compartmentation

Tear faults (or **transfer faults**) accommodate differential displacement of different parts of a **segmented** thrust sheet. If these faults are inclined, they form lateral ramps for the moving thrust sheets.



Exercise
Draw transfer faults involving folding on one side or two segment of thrust sheet.

Local thrusts

Small compression zones occur in relation to local structures.

Thrusts related to folds

When folding can no longer absorb imposed shortening (for example limbs cannot be rotated any closer together), thrust faults cut the steep or overturned limb (fold-propagation fault).

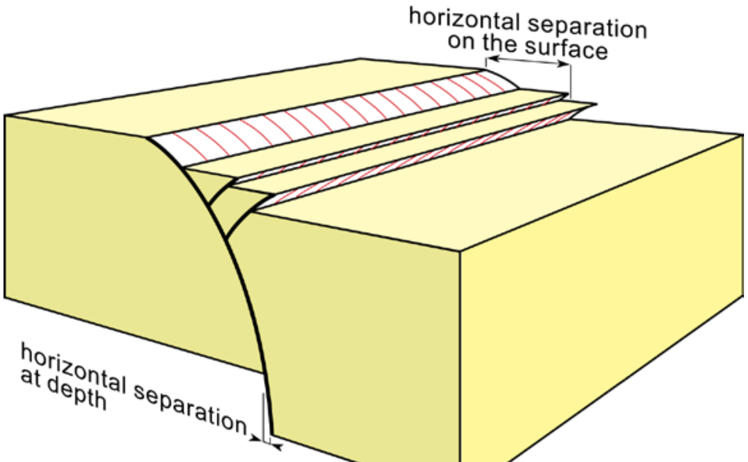
Thrusts related to domes

Diapirs and many dome structures are due to material (e.g. salt and plutonic domes) moving up through denser rocks. The raising material may push the surrounding rocks out of the way, hence forcing peripheral shortening-zones and triggering limited thrusting.

Thrusts related to normal faults

Upward-flattening normal faults

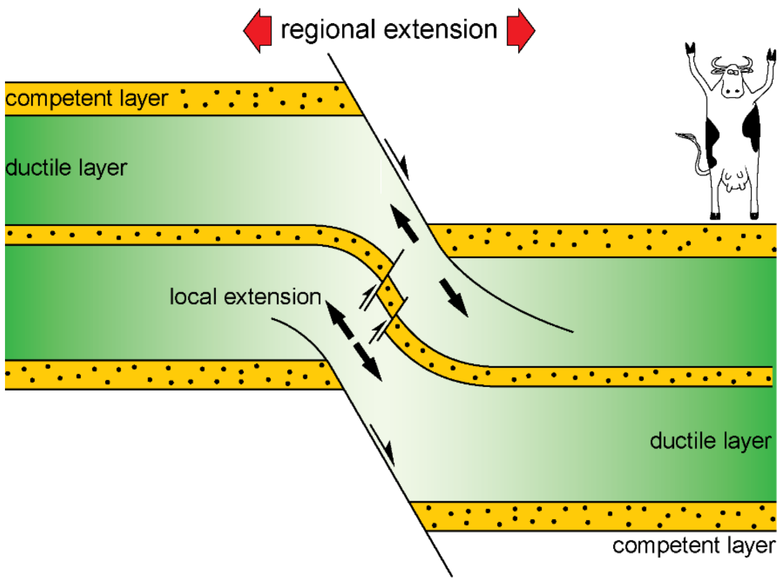
Local compression (shortening) in the hanging-wall of a concave-downward (anti-listric) normal fault produces near-surface thrust faults because there cannot be any hole between the hanging wall and the footwall. These thrusts are sub-parallel to the main normal fault.



Reverse subsidiary faults
in the hanging wall of a convex-upward normal fault

Strain accommodation thrusts in major normal faults

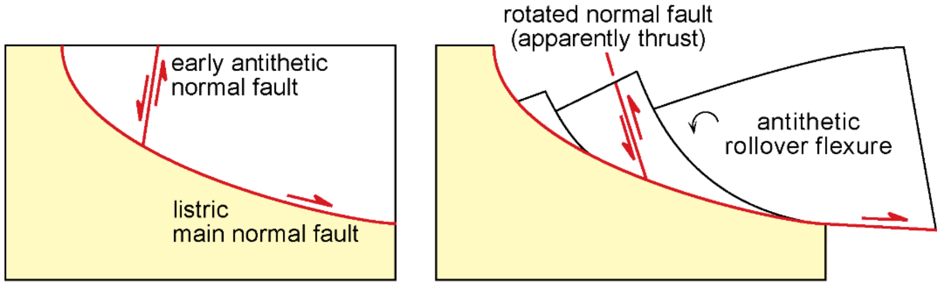
Reverse faults may form in tilted layers to accommodate layer-parallel stretching due to larger scale normal faulting.



Reverse fault accommodating local, layer-parallel extension
in a regional extension system

Rollover

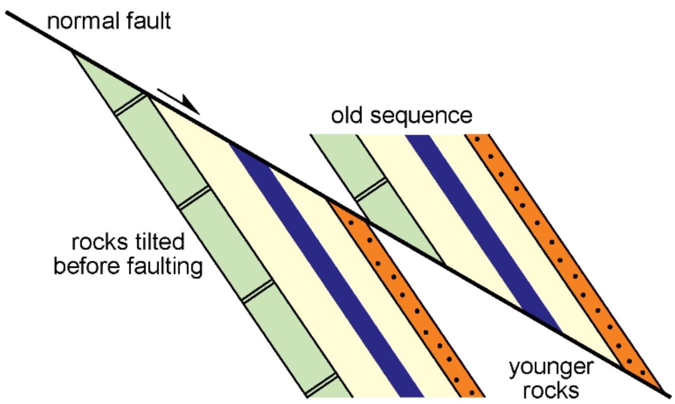
Faulting in a roll-over anticline may produce a small number of reverse secondary faults, although the main deformation is extensional. Reverse faults may be present both as direct result of this genetic process and as a result of later rotation.



Back-rotation of an antithetic normal fault to the attitude of
an apparent reverse fault in the hanging wall of a listric normal fault

Normal fault duplicating rock sequences

A normal fault cutting a previously inclined sequence may bring older on younger rocks.



Normal fault thrust-like duplicating a rock sequence

Gravity tectonics

Gravity gliding models propose that thrust sheets move down a plane inclined towards the foreland under the action of gravity (like downslope landslides or olistostrome emplacement). The sole thrust in front of the allochthonous thrust sheet may re-surface in the hinterland as an extensional feature. Thrusts and folds occur in the frontal area of the allochthonous sheet that has slipped towards the foreland once it has become gravitatively unstable. Those are usually shallow fault systems (*see further down in this lecture*).

RHEOLOGICAL CONTROL OF THRUST SYSTEMS

There are two schools of thought related to thrust tectonics:

- One is that major thrusts flatten at depth to join with some decoupling horizon, which gradually works its way back by some staircase trajectory to the original source of thrust movement (eventually, subduction plane).
- The other is that thrusts are steeper at depth, presumably to die out in ductile strains in the metamorphic lower crust or the mantle.

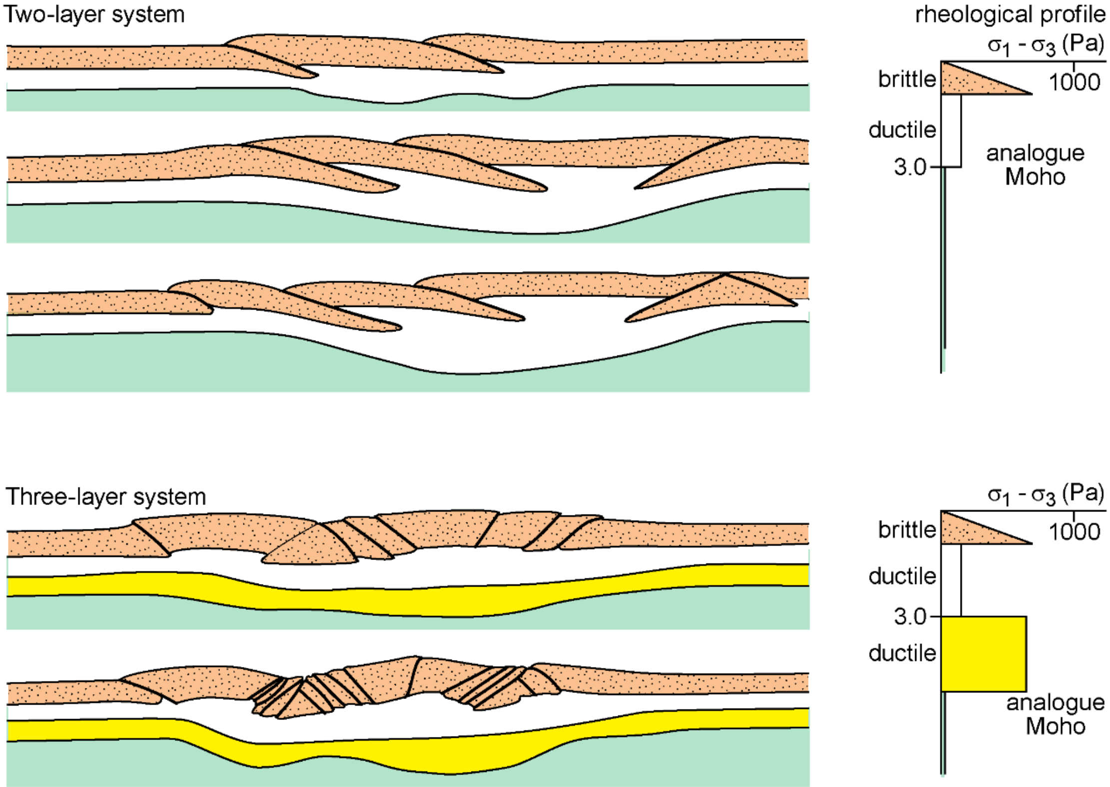
In this discussion, two parameters exert a strong influence on the deformation pattern: (1) the rheological layering and (2) coupling between brittle and viscous layers. Both parameters control whether a decoupling horizon dominates and accommodates the tectonic shortening.

Effects of the rheological layering

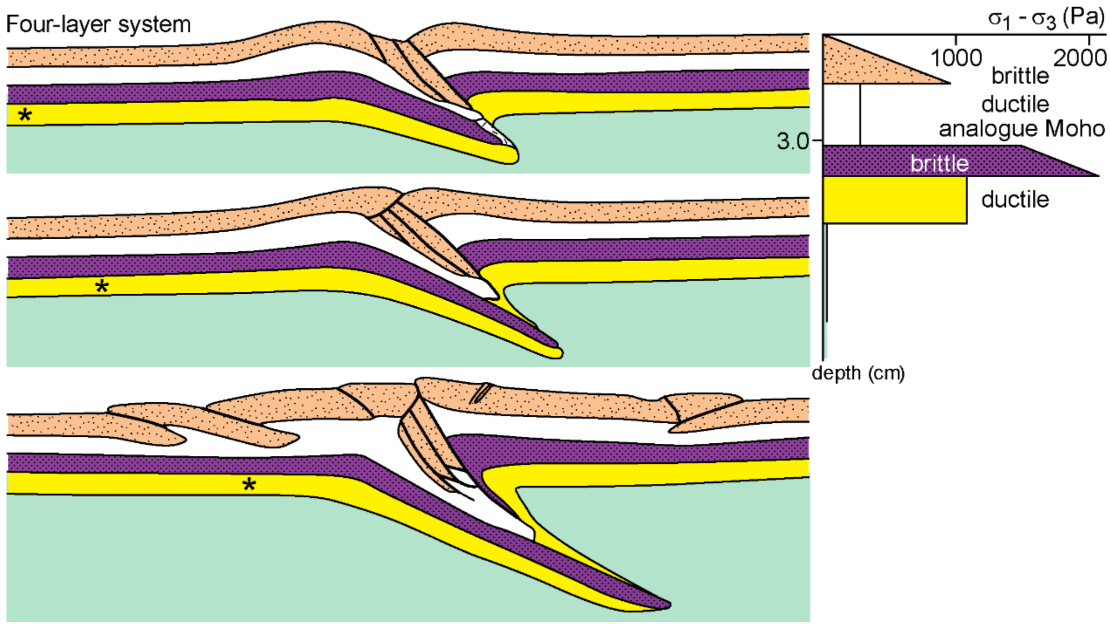
Analogue models have suggested the following large-scale behaviours:

- Two-layer (brittle/viscous) and three-layer (brittle/viscous/viscous) systems produce wide zones of distributed shortening, with conjugate thrusts in the brittle layer. The deformation zone widens with increasing shortening. Such models do not apply to modern convergence mountains.

Influence of the rheological layering on the mode of shortening in sand-silicone analogue models
after Brun (2002) *Geol. Soc. London. Spec. Pub.* **200**, 355-370

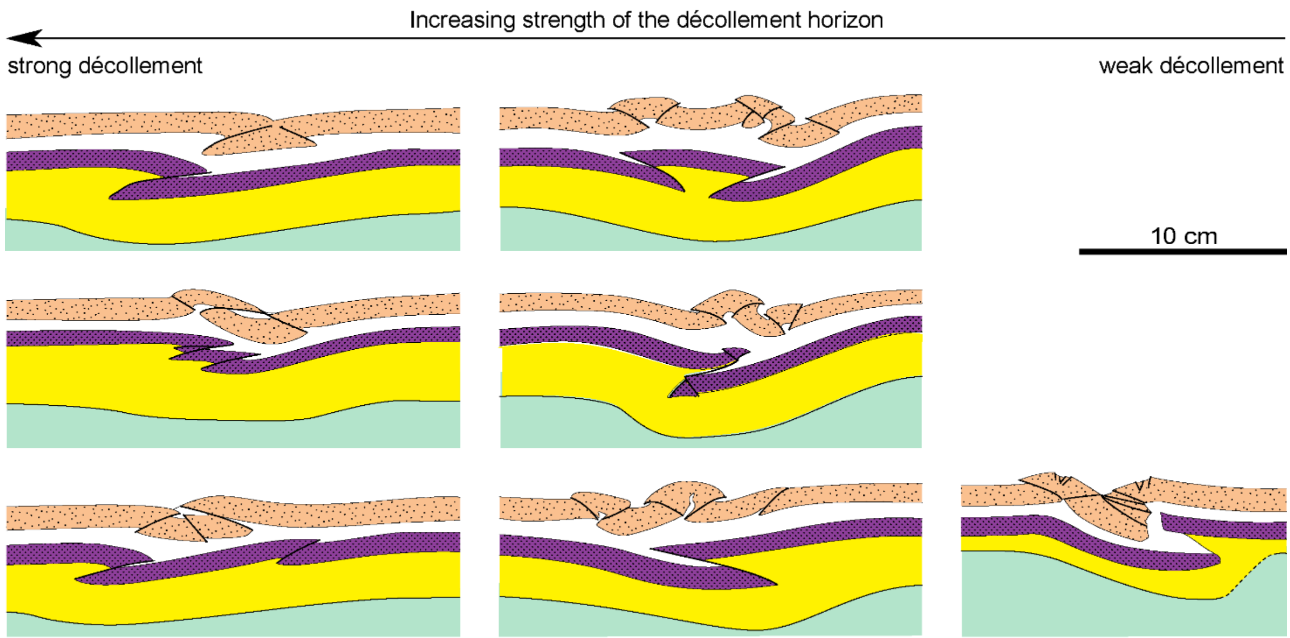


- Four-layer (brittle/viscous/brittle/viscous) models result in efficient decoupling within the highest viscous layer, which acts as a décollement level. The upper brittle layer adopts its own style of deformation, with pop-up and pop-down structures independent from thrusts in the lower brittle layer, which have a variable vergence and a larger spacing. If coupling is strong, then the asymmetry of upper layer deformation reflects the thrusting asymmetry in the lower layer.



Effects of the décollement strength

Models show that the strength of the décollement layer affects the development of passive versus active roof duplexes, the amount of translation on the individual thrusts and the ramp spacing. The presence of relatively strong décollements promotes local underthrusting of the cover, individual ramp-anticlines, internal deformation of thrust sheets, low early layer-parallel shortening, and in sequence propagation of structures.



Parallel sections through four-layer analogue models deformed in compression
after Brun J.-P. (2002) *Geol. Soc. London Spec. Pub.* **200**, 355-370

Weak décollements promote fore-thrusting of the cover, antiformal stacks, coeval growth of structures, and low internal strain, with the exception of significant early layer-parallel shortening. The strength may change along the décollement surface (for example where a salt layer stops). The

strong part may act as a pinning frontal buttress that hinders forward propagation. Shortening displacement then favors the formation of out-of-sequence thrusts and backthrusts.
Such experiments demonstrate that the mantle rheology strongly controls the first-order structures of mountain systems.

Sand box and critical taper theory

The mechanical development of folds-and-thrusts-belts is compared to the piling up of loose sand in front of a bulldozer as it is pushed up a slope. The Swiss analogy would be the behaviour of snow ahead of a snowplough. The sand (or the snow) forms a wedge shape immediately. As the bulldozer moves on, the wedge widens and increases in volume while its upper slope steepens or flattens until the frontal angle of the wedge reaches a value called the **critical taper.** At this point, the wedge is in dynamic equilibrium. It moves stably along its base, and is at the point of failure throughout.
In the critically tapered, stable wedge, equilibrium between three main elements exists:

- Frictional resistance to sliding along the base, which refers to the basal traction of the wedge.
- Forces pushing at the rear of the wedge, which express the regional tectonics.
- The shape of the wedge, which is controlled by various factors such as frontal or basal accretion, internal deformation, sedimentation, surface and tectonic erosion.

A change of one or more of these factors generates within-wedge deformation caused by internal stress release to regain or maintain stability.

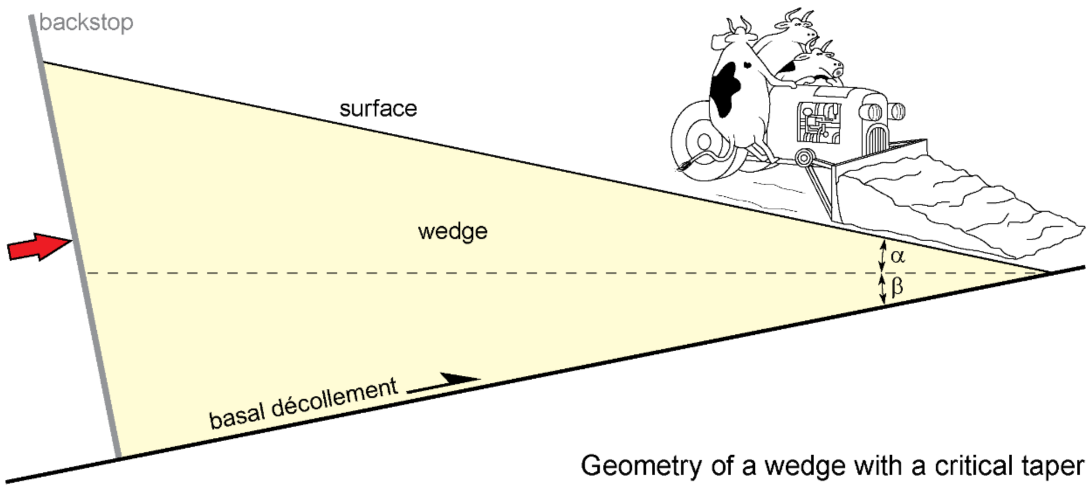
Internal forces and stresses

The gravitational potential energy, due to elevation of the hinterland, creates both horizontal and vertical stresses. If the surface of the wedge becomes too steep because of excessive thickening, then the basal plane is unable to support the load and the wedge collapses forward. If the surface of the wedge is too gentle, then not enough gravitational force is transmitted down to the basal plane to allow slip to occur and the topography builds up while the wedge deforms internally by forming folds, faults and penetrative strain.
An increase in the sliding resistance increases the critical taper, since it is the drag on the base that is fundamentally responsible for the deformation. An increase in the wedge strength, on the other hand, decreases the critical taper, since a stronger wedge can be thinner and still slide over a rough base without deforming.

Shape of the wedge

The shape of the taper is defined by the angle θ, which is the sum of the upper surface slope α, down to the foreland, and the dip of the décollement β (or basal slope), towards the hinterland.

$$\theta = \alpha + \beta$$



Geometry of a wedge with a critical taper

If material is added to the wedge so as to increase the taper above this angle, gravity spreading will reduce it. Conversely, if the wedge extends so that the taper is reduced below the critical value, stress applied to the back edge will shorten it until the critical taper is achieved.
The equation relating all the various quantities for a subaerial wedge is:

$$\alpha = \left[(1-\lambda_b)\mu_b - (1-\lambda_i)k\beta\right] / \left[(1-\lambda_i)k + 1\right] \qquad (6)$$

where the strength (yield stress) of the rock in the wedge is k.

λ_i is the ratio of pore fluid pressure to overburden internal to the wedge.

λ_b is the ratio of pore fluid pressure to overburden along the basal décollement.

μ_b is the décollement friction.

Geological application

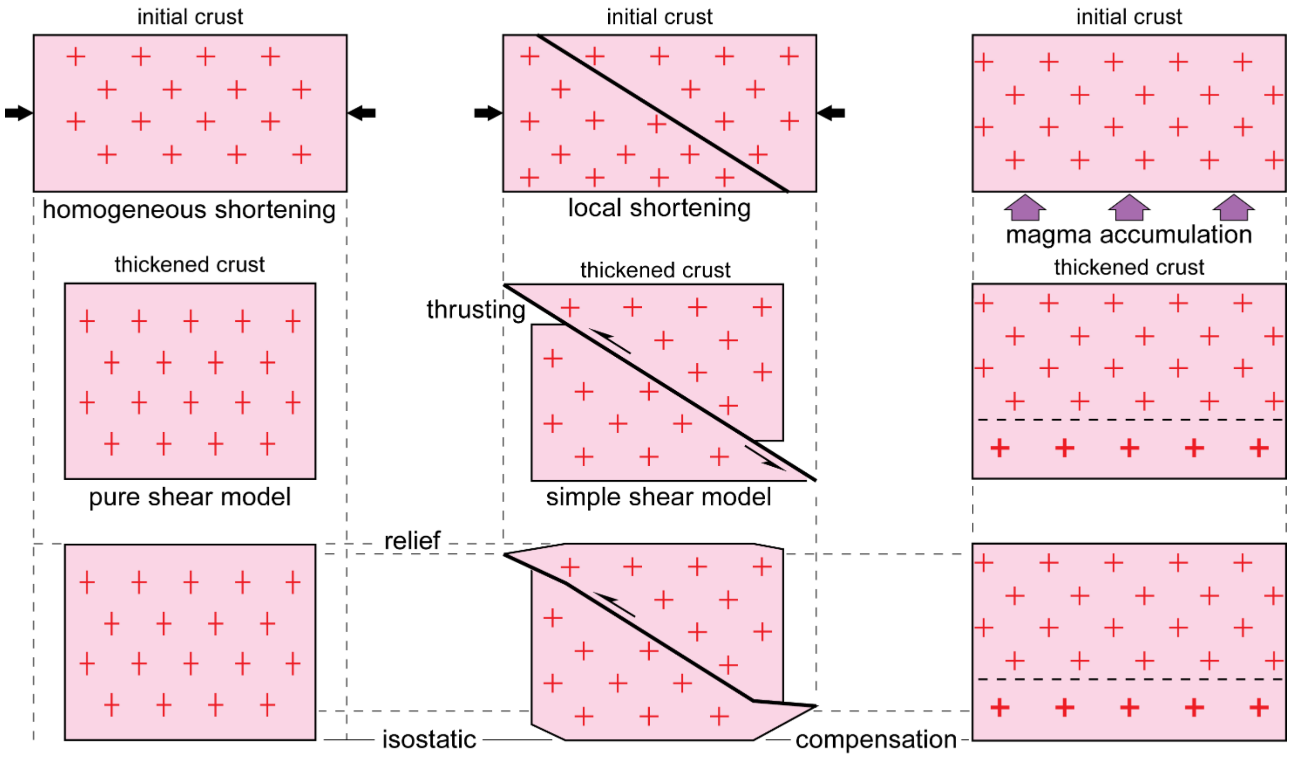
A geological wedge is believed to evolve by the addition of sediment scrapped at its toe from the down-going slab. In support, seismic profiles across convergent mountain systems frequently show that a major décollement separates the colliding plates. Large amounts of subhorizontal motion take place on this décollement, also known as **sole thrust** (or **basal thrust**), which dips gently towards the overriding plate. Motion along the décollement results in the tectonic accretion of imbricate slices one on top of the next in the deforming hanging wall through distributed horizontal shortening as well as folding and faulting.

- The hanging wall develops into a wedge-shaped tectonic unit, when viewed in cross-section parallel to the movement direction, with the narrow end in the direction of motion.
- The subducting footwall, however, remains relatively undeformed, which typifies a thin-skinned thrust belt.

Moving thrust wedges are driven by plate convergence but the geometry acts in response to the rate of convergence and to the strength of the basal detachment. Hence, they are in a state of dynamic equilibrium. The sole thrust is considered to be weak while the wedge material follows the Mohr-Coulomb failure criterion. The corresponding Mohr construction implies that rocks effectively increase in strength as lithostatic pressure rises; hence the rocks at the back end of the wedge are effectively stronger than those at the front. Variants of wedge models have incorporated a number of different types of rheology.
The wedge models link the topography of orogenic belts to the rheology of the crust and includes the effects of body forces and externally applied tectonic forces. The theory is developed in another lecture.

LARGE-SCALE ANALYSIS OF THRUST SYSTEMS

Mountains resulting from horizontal space reduction are among the most attractive structures on the Earth's surface. In map view, they form long belts occasionally curved in **oroclines**. Sinuosities are likely inherited from irregularities of the plate margins before orogeny.



Three mechanisms of crustal (lithospheric) thickening

General features

Maps and profiles show that:

- Thrust systems are long and relatively narrow.
- Deformation progressively **migrates** towards the foreland.
- Deformation ends in the foreland at a **front thrust**.

Cross sections of thrust systems, whether thin- or thick-skinned, have a general wedge-shape thinning towards the foreland. This geometry has led to the concept of **orogenic wedge**, which considers that thrust-belt mechanics is analogous to pushing sand or snow uphill in front of a moving bulldozer.
In reality, compressional mountains have two sides, often corresponding to two asymmetrical orogenic wedges sharing the same highest elevation axial ranges. One side, usually the wider, is the **pro-wedge**, the other side the **retro-wedge**. The structural asymmetry reflects asymmetry of the tectonic system. Pro-wedges are mostly developed on the subducting plate, with thrusting synthetic to the subduction direction. Retro-wedges mostly form on the overriding plate, with backthrusting antithetic to the subduction direction. The general fanning configuration is **bivergent**.

Plate coupling – high stress / low stress convergent plate boundaries

Depending on the nature of the colliding plates and on the efficiency of slab pull, the dynamics of plate convergence is categorized into two end-members, which refer to the forces at the contact between the two plates.
If the two plates are **coupled**, the undergoing plate pushes under the hanging wall plate. The plate boundary is then under high compressional stresses, which can be transmitted within the two lithospheres. This can be envisioned where the buoyancy of the descending plate (e.g. carrying continent, island arc, oceanic plateau, spreading ridge) resists subduction.
If the two plates are **decoupled**, the undergoing plate tends to separate from the hanging wall plate. The plate boundary is then under low stresses. Suction forces may even pull the hanging wall plate which is then under tension. This can be envisioned where slab pull is important, often triggering trench roll back.

Thermal effects

Large scale shortening / thickening of the crust modifies the thermal gradient, principally by conduction and advection: Footwall rocks carry down their temperature while they are loaded by thrust sheets. The thermal gradient decreases since relatively cold conditions are taken deeper in the crust. Radiogenic heat plays a secondary role in these considerations. Consequences are that metamorphic "high pressure" conditions result from thickening. Re-equilibration towards normal thermal gradient will correspond to reheating buried rocks. At that stage melting of root zones generates magmatism; plutonism is an important heat advection process in waning stages of collisional orogens. It is even more important and long lasting process in arc systems.

Thrust systems

Thrusting is a primary mountain-building mechanism. The modern classification of mountains and their thrust systems refers to their context in a plate tectonics framework. As such, four types of mountains are distinguished, which may actually represent four stages of convergence cycles. They are:

- Subduction mountains, subdivided into "cordilleran" mountains where the magmatic arc is installed on a continent (Andes), and "insular" mountains where the arc is on oceanic lithosphere (Indonesia).
- Obduction mountains, where the oceanic lithosphere is thrust over the continental one (Oman).
- Collision mountains involving the initial contact and the development of a suture zone between two continental lithospheres after resorption of oceanic lithosphere. A modern example is the collision between the northwestern Australian continental margin and the Banda arc in the Timor island region.
- Intracontinental mountains that form within continental plates, away from any plate boundary (Pyrenees, present-day Tien-Shan, Atlas, Himalayas). Continued convergence is accommodated along the suture zone, and in intraplate, crustal-scale thrusts. Intense mountain building processes and the development of high plateaus characterize this stage.

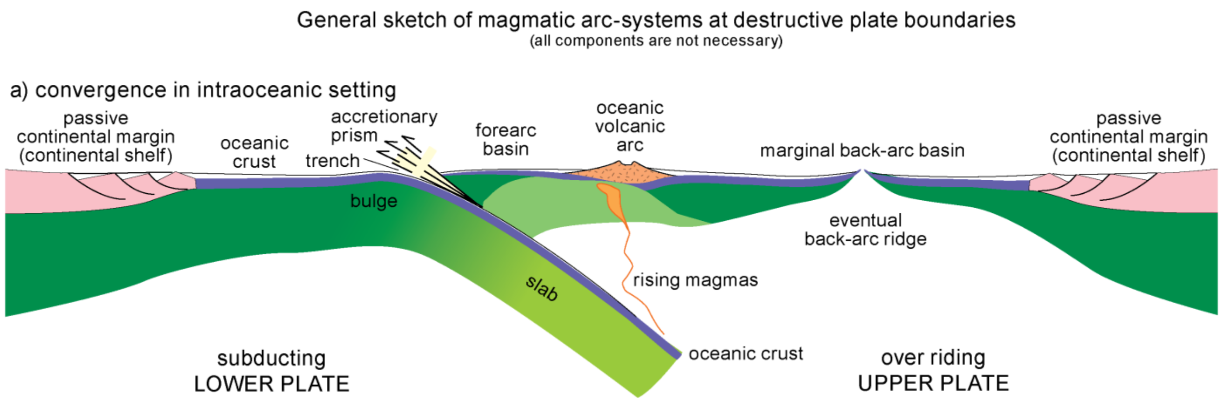
When convergence ceases, erosion and isostatic adjustment prevail to expose the roots (Variscides). The whole evolution may last tens of millions of years and possibly longer. Juxtaposed in time, these stages can be superposed in a single mountain system that had a long evolution. Therefore, ancient tectonic reconstructions rely upon the identification of rock assemblages characterising plate boundaries and the large-scale geometry of the thrust system that was built.
Three systems with different form of the sole thrust in the hinterland are geometrically complete.

1. Subduction systems with no deformation of the basement are typically pre-collision features. The sole thrust dips into the subduction zone and may record the bulk plate tectonics movement. Subduction systems do not require a root-zone as the source area for the thrust sheets identified in other fold-and-thrust systems. There are about 50 000km of convergent plate margins in the world.
2. In collision belts, the sole thrust of the fold-and-thrust belt dips in the **root zone** under the metamorphic rocks of the hinterland. Compression is transmitted by the hinterland (which has shortened under different modes) to the foreland. This system is common in orogenic belts and may accommodate large amounts of shortening – displacement.
3. The trailing edge of the sole thrust cuts up-section towards the surface so that the shortening along splay thrusts in the frontal area is balanced by extension along imbricate listric normal faults in the hinterland region. Paired shortened and lengthened belts usually imply moderate displacements and are gravity driven systems of thrust sheets gliding down slope away from the orogenic elevated interior. The process refers to post-collision gravity sliding and spreading.

Intra-oceanic subduction: Island arc systems

The descent of one plate beneath the other is the common response to the space problem posed by convergence. This response is particularly common when an oceanic lithosphere collides with a continental one. The denser oceanic plate is forced down into the mantle beneath the more buoyant continental plate. Similarly, if two oceanic lithospheres collide, the denser (likely the older) plate flexes and sinks into the asthenosphere beneath the other plate. This mechanism is called **subduction**. Since subduction involves the consumption of a plate into the Earth's interior, subduction zones are also termed **destructive plate boundaries**. A subducted plate is a **slab**. Slabs descend into the asthenosphere with an average dip angle of about 45° but, depending on buoyancy, this angle may vary between less than 10 and 90°. Subduction entrains seawater and probably small amounts of sediments at mantle depth. Since cold crustal material is subducted to great depths in a relatively short time, isotherms are buckled down, which leads to **high-pressure/low temperature metamorphism** in the subduction zone. At a depth of 100 to 150 km, dehydration of the subducted oceanic crust starts, and fluids raise from the slab into the overlying mantle. Heating the slab and hydrating a portion of the mantle causes extensive mineralogical changes and partial melting. Partial melting of the down-going slab, the overlying mantle wedge and the basal continental crust generates magmas. The contribution of each possible source influences the composition of the resulting igneous rocks but tholeiitic and calc-alkaline rocks dominate all variants. Magmas rise toward the surface, eventually make their way up into the leading edge of the overriding plate, where they add material to the crust and build volcanoes above it. If the upper plate is oceanic, the volcanoes pile up until they poke through the surface of the ocean. The general consequence is a nearly systematic association between subduction and magmatic activity.

Most subduction zones at the present time are situated at volcanic **island arcs** within the oceans. The term arc in this denomination refers to the convexity towards the subducting plate in map view. This convexity is due to the spherical geometry of the plates. **Intra-oceanic** subduction zones comprise four important main components with characteristic morphology and characteristic rock associations. A systematic arrangement of these tectonic elements provides a convenient reference framework for comparison among arcs, knowing that all elements are not present within every island arc system. The recognition of ancient arcs and their polarity is critical to the reconstruction of past tectonics because of their consistent spatial association with down-going slabs.



Island arc

The island arc consists of partially submerged, volcanic mountain ranges that occur on the overriding plate, 60 to 170 km above the top of the slab. This relationship would assign some systematic role to slab dip and convergence rate to arc formation and location. Metamorphic dehydration reactions take place in the down-going slab, and the influx of released volatiles triggers partial melting of the overlying mantle wedge. The mantle wedge actually is the principal site of magma development. Under the influence of gravity, such high-temperatures, low-density magmas buoyantly rise into and through the overriding plate. The products of intrusion and extrusion contribute to the formation of a **magmatic arc** parallel to the convergent plate boundary. The uppermost parts of the magmatic arc comprise a **volcanic arc**. Reference examples of island arc systems border the Pacific Ocean. Typically, calc-alkaline basalts and andesites predominate, while dacites and rhyolites are relatively rare. The series consists of silica-oversaturated rocks that tend to contain more Al_2O_3 than tholeiitic lavas, and their intermediate members do not normally show the effects of significant Fe-enrichment. Boninite, a Mg-rich and Ti-poor lava, is unique to these arcs. Plutonic rocks are typically I-type gabbros and diorites, with subordinate plagiogranites. Continental crust is not necessarily involved (Aleutians), but some island arcs are formed from pieces of continental crust (Japan) that have separated from a nearby continent.

Oceanic trench

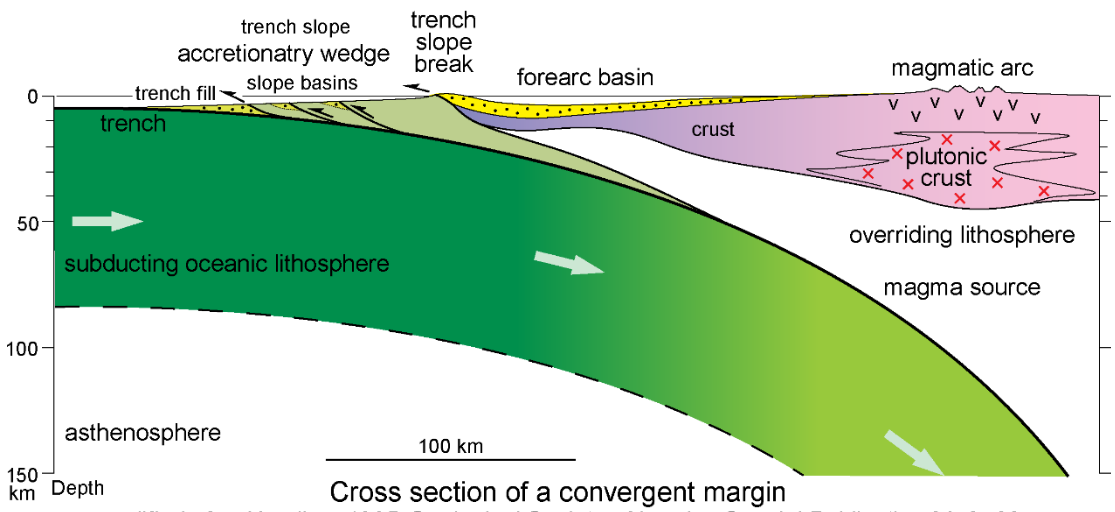
The **oceanic trench** is a several hundred kilometres long and narrow topographic depression of the sea floor. It traces at the Earth's surface, on the convex side of the island arcs, the boundary between the downgoing and overriding plates. The trench marks the position at which the flexed slab begins to go under. Trenches are deep because the slab pulls downward the plate (water depths of c. 10 km with a breadth of c. 100 km, e.g. the Mariana and Kurile trenches). Therefore, trenches are important sites of sedimentation (**trench fill**), dominantly turbiditic with minor pelagic components. Seismic investigations across ocean-trenches show a typical, asymmetric V-shape with the steeper (10-15°) side facing the plate that is being subducted. This "wall" marks the edge of the overriding plate and the outermost fore arc.

Bulge

The elastic response of the subducting plate as it bends to descend into the mantle probably causes the approximately 200 km wide and 200-400 m higher than ocean floor **bulge** or **outer swell** found on the lower plate, 100-250 km seaward from the trench. Bending of the lithosphere produces tension in the shallow crustal levels, which may lead to normal faulting. Subsequent grabens, parallel to the trench, are traps in which overlying sediments can be entrained into subduction to mantle depth. Owing to the low heat flow, metamorphism that may occur in the deep part of the sedimentary pile is of high-pressure–low-temperature type.

Forearc

The **arc-trench gap** or **forearc** is located between the arc and the trench and has a width that depends strongly upon the dip of the slab. The forearc basin comprises hemipelagic and clastic sediments derived mostly from the arc. The fore-arc basin is usually little deformed, demonstrating that the overriding plate is not affected by convergence-related shortening. In a simplistic view, the forearc plate is compared to a bulldozer blade that scrapes material from the top of the subducting plate. Consequently, the rather undeformed forearc basin may cover a thick, wedge-shaped package of highly deformed pelagic and trench-derived sediments imbricated with slices of trench and oceanic material scraped off the descending slab: the **accretionary wedge**, whose surface slopes toward the trench.



Cross section of a convergent margin
modified after Hamilton 1995 *Geological Society of London Special Publication* **81**, 3–28

Accretionary wedge

Accretionary wedges, at the front of the overriding plates, are pre-collision, broad zones (c. 100 km) of crustal deformation in a subduction zone. They exhibit many characteristics of fold-and-thrust belts and their developments are probably similar. The wedge is separated from the subducting slab by the basal décollement. The rear buttress (**backstop**) is what obstructs the horizontal movement of the wedge sediments. It can be the front of the overriding plate or material of the wedge itself. Most backstops dip towards the trench, but arcward dipping backstops are known. Deformation within the wedge accommodates the influx of material brought into the trench by the incoming, subducting plate but jammed against the backstop. Nearly half of the convergent plate boundaries worldwide possess an accretionary wedge in which incoming sediments pile up: these boundaries undergo **accretion**. The other half lacks an accretionary wedge; in some cases, this is attributed to the subducting plate slicing and dragging the frontal part of the forearc down into the mantle. Such boundaries are erosive.

Accretion

If the sediment flux is high, pelagic sediments and ocean-floor basalts of the down-going lithosphere are progressively sheared off by the leading edge of the overriding plate. Scraped off material is incorporated into the base of the accretionary wedge, a process called **tectonic underplating**. In this case the accretionary wedge grows from its base by the addition of sedimentary layers from below. The trench migrates away from the magmatic arc over the life of the convergent margin. Thrusting and associated folding denote progressive shortening and thickening of the accretionary wedge.
Characteristically, rocks in the accretionary wedge are cut by numerous imbricate thrusts that are dominantly synthetic to the subduction zone and merge into the basal décollement that separates the subducting from the overriding plate. The rocks in the deep parts of the accretionary wedge are metamorphosed in a low-temperature/high-pressure environment to produce blueschists. In some cases the deformation is so intense that stratigraphic continuity is destroyed. Such chaotic, mixed deposits, with millimetre to kilometre big sedimentary fragments and blocks of basaltic and ultramafic rocks included in a fine-grained sedimentary matrix constitute **mélanges**. Tectonic stacking of thrust sheets in the accretionary wedge builds a structural high, the **fore-arc ridge**, which bounds the forearc basin on the ocean side.

Subduction erosion

If the sediment flux is low, all incoming sedimentary material is subducted. Material tectonically eroded at the base of the forearc wedge is transferred from the overriding plate to the subducting plate and carried down the subduction zone. This process of tectonic ablation is known as **subduction erosion.** In this case the location of the trench will migrate towards the magmatic arc over the life of the convergent margin. This process can remove the entire fore-arc plate.

Back arc

Back-arc regions separate the island arc and the continent of the upper plate. The oceanic crust of the back-arc region forms an inactive **marginal basin** between the active arc and the adjacent continent (e.g. west Philippine basin). The simplest cases are trapped ocean lithospheres that reside behind an island arc (e.g. the Bering Sea behind the Aleutian Arc).
Back arc regions undergo compression or extension or strike slip deformation, depending on the plate dynamics.

Compressional back arc

Compression in back arc regions seems to depend on the subduction angle. Thin skin and thick skin tectonics may develop, depending on the response of the shortened crust.

Extensional back arc

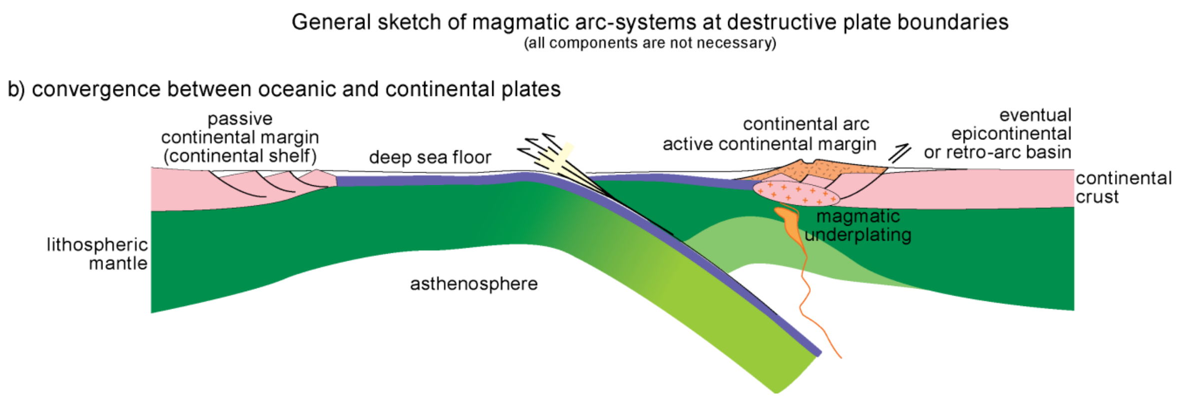
Rifting in well-developed arc systems may generate a new sea floor spreading center behind the island arc (Philippine Sea behind the Mariana Arc), presumably as a result of complex convective eddies in the asthenosphere, above the subducting plate. **Back-arc spreading**, which is extension and spreading of the sea floor behind the island arc, is similar to the seafloor spreading at ocean ridges; creation of new oceanic floor in the **back-arc basin** causes oceanwards migration of the arc and its forearc with respect to the continent. Extension is also favoured by trench roll-back. Rifting may split the arc. An **interarc basin** is then opened between the extinct **remnant arc** and the active arc (e.g. Mariana). The back-arc has an oceanic crust and abyssal depths, and may contain alkaline, shoshonitic magmatism is developed (Japan Sea). Depending on proximity to the arc, sediments are volcanoclastic, hemipelagic or pelagic.

Subduction at continental margins

Where oceanic and continental plates converge, the less-dense continental crust resists subduction into the mantle and overrides the oceanic plate (the Andes, along the west side of South America, for example). The subduction zone is **subcontinental**.

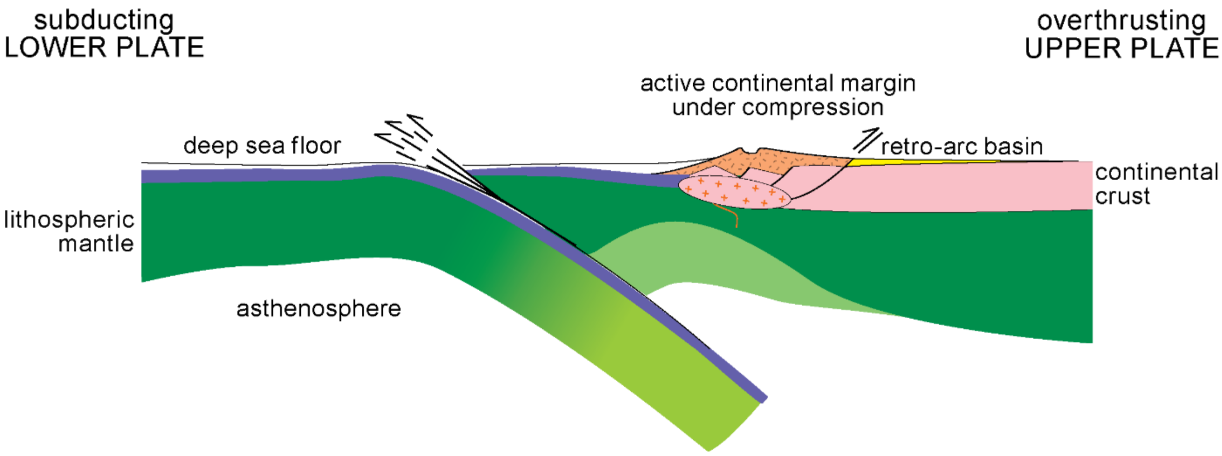
Magmatic arc

Like for intra-oceanic subduction, slab melting at a depth of 100 - 150 km generates magma, which is less dense than the surrounding mantle. The magma rises into the overriding continental lithosphere and may melt and incorporate some silica-enriched crustal rocks. The subsequent calc-alkaline magmatic arc grows on the continental plate. The complete suite of igneous rocks includes andesite- and dacite-dominated lavas, with subsidiary basalts and rhyolites, along with granodiorite- and tonalite-dominated plutons, with a minor amount of gabbro, diorite, and granite. Large plutonic bodies coalesce and form a long and linear **batholith**. The silica-rich magma partly erupts in explosive and dangerous volcanoes (e.g., Mount St. Helens in the USA, or Mount Menapi in Java, Indonesia). The tectonic system is called an **active continental margin**.



Back arc

In continental arcs, the back-arc region may be either exposed or inundated to form a shallow marine region. Its structural evolution depends on whether it is under extension or compression. The dip-angle of the slab, which affects coupling between the overriding and the subducting plate, is for that essential. Extension, which produces episodic formation of back-arc and/or marginal basins, is often related to steep slabs and roll back. Compression is often related to locking of the subduction and/or from the low dip of the subducting slab below the continental upper plate. In these cases strong coupling between the two plates generates compression. Compressional margins possess a thickened crust and high mountains in the axial part of a **bivergent** orogen, i.e. thrusts with opposed movement directions border the mountain belt. Thrusts move away from the arc and stack parts of it on the continental crust, which loads the plate and causes subsidence. Seaward verging thrusts follow the polarity of the subduction zone, whereas the opposite craton-ward thrusts produce **retro-arc basins** in which coarse, terrigenous sediments are common filling materials.



Collision

If the subducting plate carries a continent, the buoyancy of the continental lithosphere entering the trench restrains subduction. If both plates carry continents, the closing ocean brings inevitably its former continental margins together. The oceanic slab tends to pull down the continental margin to which it is attached. But the two continental lithospheres are usually equally buoyant and neither of them will easily be driven beneath the other. This confrontation leads to increasing horizontal compression until subduction is locked. Regional-scale shortening then occurs, producing a zone of very complex structure where folding and thrusting are intimately associated. The two continental masses are ultimately welded together into a single continental block. **Collision** describes this response to plate convergence. Collision forms a new mountain range characterized by high elevations and a continental crust that may reach more than twice its normal thickness. This new, orogenic crust is a tectonic assemblage of accreted oceanic lithosphere, magmatic arcs, and continental margins with associated sediments.
Three types of collision are distinguished:

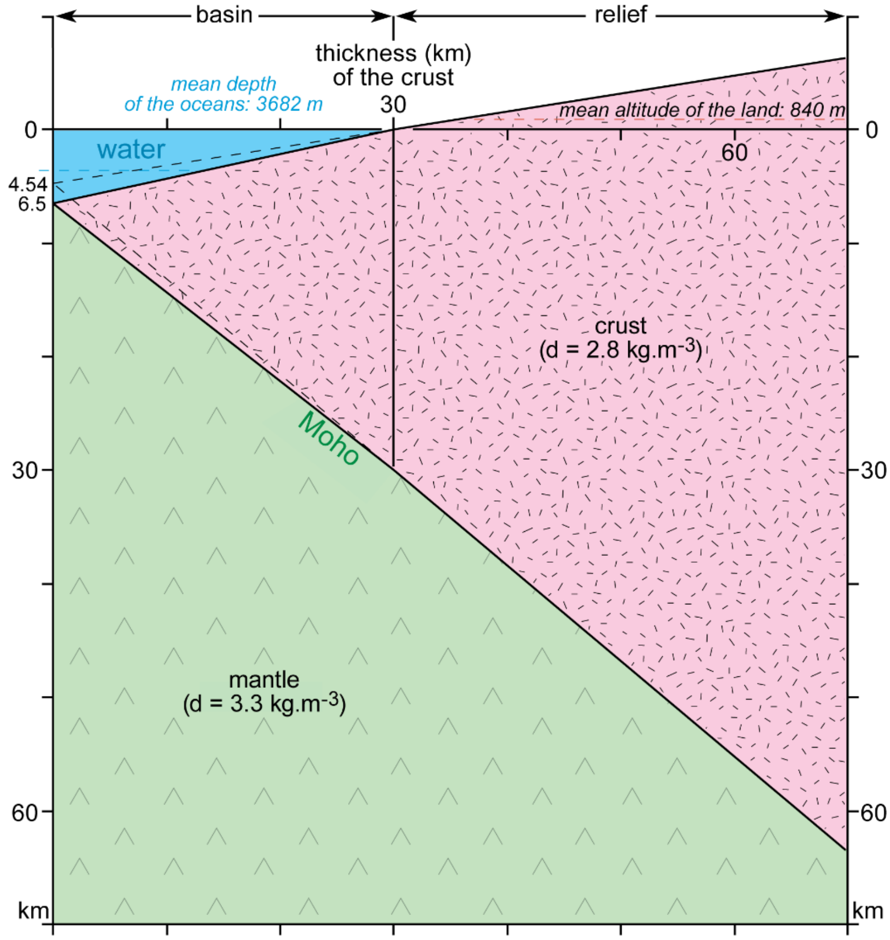
- island arc vs. island arc: A recent incipient collision of this type is inferred in the Molucca Sea.
- continent vs. island arc: the most conspicuous example is convergence between the Banda islands and the Australian continent.
- continent vs. continent: the reference example is collision between India and Asia.

If convergence continues after collision, subduction may take place behind the continent or arc **docked** to the main continental mass. If the docked plates are small, the plate tectonic system remains roughly similar and no change in plate motions is required. If the new subduction is shifted far from the collision site, new plate boundaries are then created and a global rearrangement of plate tectonics takes place.
Collision orogens display a large structural diversity. However, whatever their size, they share some important characteristics.

Thickened crust

Shortening is intimately associated with thickening because, for sake of simplicity, geological deformation preserves volumes of continental crust. Structural studies have shown that a plate thickens vertically as much as it shortens horizontally. As a corollary, the mountain belt due to continental convergence represents also a belt of thickened crust and, likely, of thickened lithosphere.
Plate tectonics tells that the crust floats on the earth mantle just as an iceberg floats on the sea, the biggest volume compared to a root remaining immerged. From the density difference between crust and mantle rocks, we know that it requires 5 to 7 km of crustal root to balance each km of mountain range above sea level. A mountain grows 5 to 7 times more downward than upward. Therefore, collisions systems are site of intense metamorphism and igneous activity.
An overthickened, continental crust tends to rise as a consequence of positive buoyancy forces, thus generating topographic elevations, a folded **collisional mountain belt**. For instance the Alpine-

Himalayan belt represents the collision of the Eurasian continent to the north with the African and Indian continents to the south of the closed plate boundary.



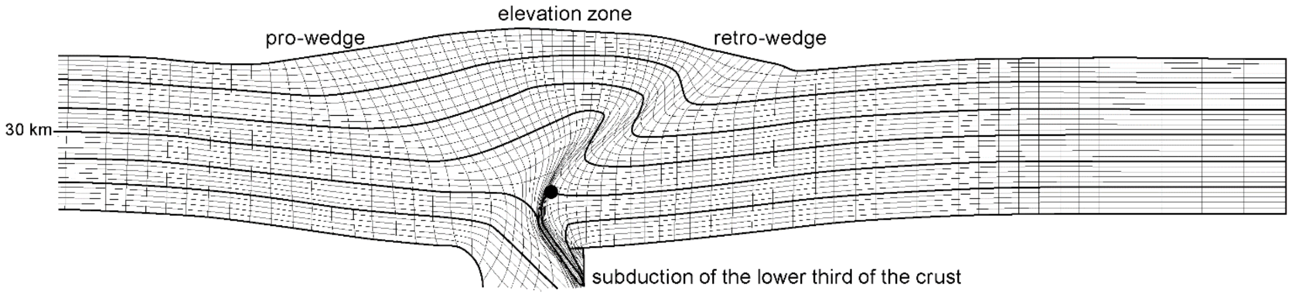
Theoretical elevation of the lithosphere surface with respect to the crustal thickness

Subducted versus decoupled crust

Thickening of the orogenic crust involves tectonic imbrication of large thrust sheets. The amount of displacement on major thrusts questions where the lower crust and the mantle of these nappes are. Two structural end-members are recognized: (1) Continental subduction, i.e. the underthrust crust remains attached to its mantle and (2) continental decoupling / delamination, i.e. the nappes of continental and/or oceanic origin are scraped off from their underlying crustal and mantle layers in a “thin-skinned” manner. The response of the lower lithosphere to collisional thickening is an unsolved question partly discussed in the lecture on tectonic systems.

Shape and asymmetry

Analogue and numerical models have shown that the length and the map form of collision orogens reflect the shape of the contact zone between the colliding plates, the amount of shortening and the strength of the lithospheres. In profile, hinterlands are thickest and the mountain system has a wedge shape becoming progressively thinner toward the foreland. Models also show that the style of thickening depends essentially on the behaviour of the upper mantle: strong lithospheres (as with a brittle mantle) tend to build narrow, asymmetric and high orogens; weak lithospheres deform more symmetrically and homogeneously and produce wide and lower collision mountains. The major thrust systems reflect the polarity direction of subduction, at least during the first stages of collision. The lower plate continental margin, which is attached to the sinking slab, is thrust below the active upper plate margin. Further structural developments depend on numerous factors that control the geometry, rate of shortening and thermal structure of the collisional zone. Many collision orogens are in fact double-vergent and asymmetric (e.g. the Alps and the Pyrenees). The well-developed **pro-wedge** is synthetic to the main subduction and deforms mostly the lower plate. During progressive collision, the pro-wedge grows by **footwall-propagating thrusting**, i.e. by accretion of crustal imbricates peeled off the underthrust lower plate at the front of the prograding (widening) orogenic wedge. The narrower **retro-wedge** affects the upper plate with a vergence opposite to that of the pro-wedge. The retro-wedge acts as the back-stop of critical wedges. Both the pro- and retro-wedge constitute the **orogenic wedge.**



2-D Finite Element Model of deformation in a collision orogen
after Beaumont C. & Quinlan G. (1994) *Geophysical Journal International* **116**, 754-783

Structural units

The most spectacular compressional effects are exhibited in the thrust and fold belt that have absorbed hundreds of kilometres shortening by overlapping crustal slices and **accretion** (i.e. tectonic assembling) of rocks from one plate onto the leading edge of the other. Each collision belt has its own character. However, a few elements are generic:

Suture and ophiolites

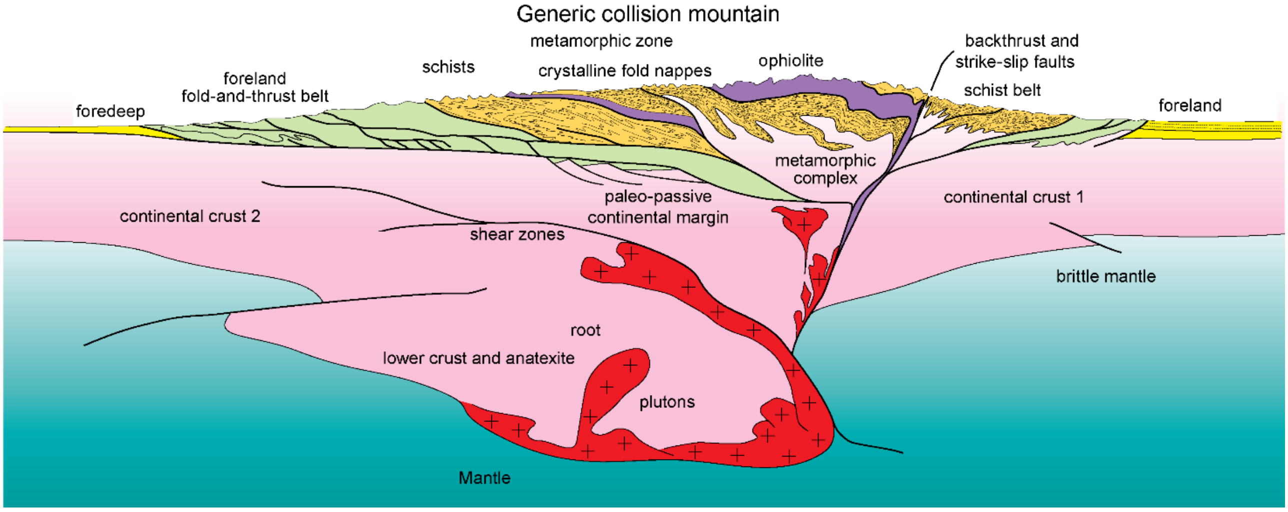
The **suture** or **suture zone** is the contact between plates that have collided and shortened. Along the suture zone, remnants of volcanic arcs and possibly slivers and obducted klippen of oceanic lithosphere from the closed ocean (**ophiolites**) are preserved. Suture zones largely consist in folded and metamorphosed sedimentary rocks. Many old collision orogens have successively involved subduction, obduction and continent-continent collision. Consequently, arc-related rocks found on one side of the suture derive from the overriding plate. Rocks on the other side of the suture are chiefly derived from the passive margin of the subducted continent. The suture is the main dividing element of collision orogens.

Metamorphic hinterland

The **metamorphic core** or axis is a very complex zone often adjacent to the suture. Because crustal rocks are buried to higher temperature and pressure conditions, the deep roots of the mountains are metamorphosed and even molten (which is called **anatexis**). Folding and thrusting contemporaneous with metamorphic recrystallisation are pervasive, dominantly in the continent being overridden. Crustal units may be transported hundreds of kilometres as **allochthonous nappes** over thermally weakened, ductile shear zones. Overlapping crustal slices and **accretion** (here meaning tectonic assembling) of rocks from one plate onto the leading edge of the other are commonly associated with intermediate-pressure type metamorphism and subsequent plutonism. While one continent is shoved above the other one, the crust is submitted to a horizontal force couple that imparts a strong asymmetry (**vergence**) of structures and resulting mountain system. The shear motion is dominantly synthetic to the continental subduction. Crustal melting generates magmas of granitic compositions that rise up into the upper crust. Peraluminous **S-type granites** may be a hallmark of collision belts. This zone in which basement, crustal rocks are intensely deformed is also termed **hinterland**. Major thrusts separate the hinterland from the **forelands**, along margins of the metamorphic axis, where the little deformed basement remained relatively rigid.

Fold-and-thrust belts

Thin-skinned **fold-and-thrust belts** are generally found between the undeformed foreland and the strongly deformed hinterland of mountain belts. Foreland **fold-and-thrust** belts are formed of sediments formerly deposited on the continental margins and now stacked and thickened away from the orogenic axis. They often also involve strata of the early part of the foreland basin, which became incorporated into the orogen. In many foreland fold-and-thrust belts, the main décollement-surface remains between the strong crystalline basement and the sedimentary cover. Listric thrust faults that splay out from the subhorizontal master detachment fault delineate individual thrust sheets, or imbricates. The general geometry is similar to that of a pile of sand pushed in front of a bulldozer.

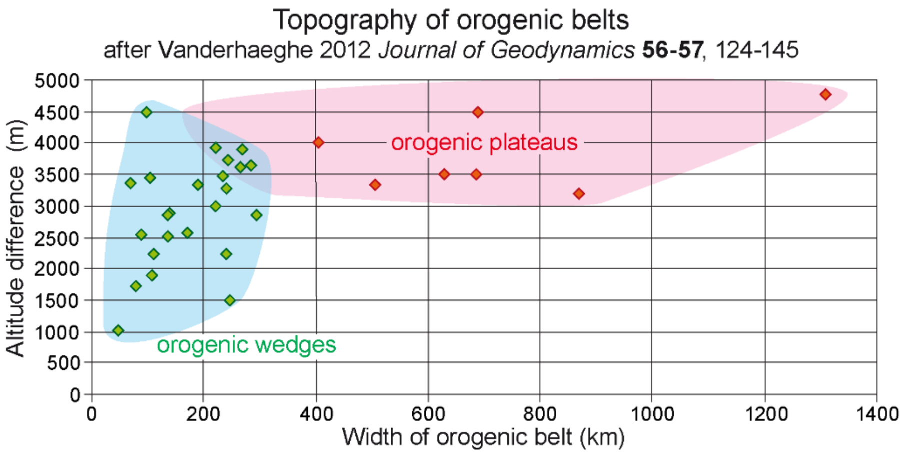


Foreland basins

Peripheral **foreland basins** result from elastic downwarp of the lithosphere under the load caused by sideways thrusting, thus material addition onto the edges of the colliding plates. Upward convex lithospheric bending in front of mountain belts forms a triangular basin (in profile) that thins out away from the mountain belt and in which material mainly eroded from the adjacent mountain belt is deposited (Molasse Basin north of the Alps, Ganges Basin south of the Himalayas). A subtle bulge may form at the external margin. The immature, clastic sediments, collectively called **molasse,** unconformably overlie older sediments and basement units and portray marine followed by non-marine alluvial fans, floodplains and lowland environments. The advance of the thrust belt pushes foreland subsidence ahead of it while early molasse sediments are progressively overridden by the orogen. With continuing convergence, new large thrusts may incorporate and carry forward the older foreland basin. Small basins passively transported and uplifted on top of a moving thrust sheet are known as a **piggy-back basin**.

Topography

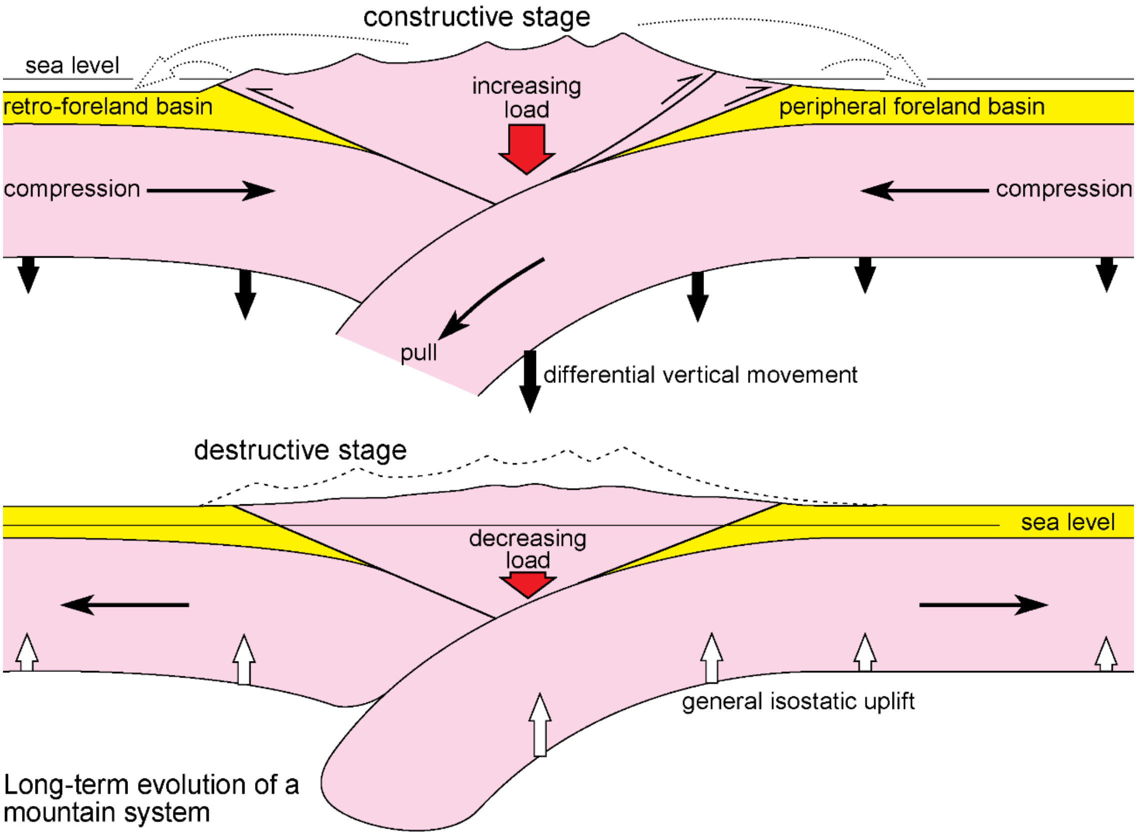
Collisional mountain belts are classified into (i) narrow ranges and (ii) wide plateaus.



The width of narrow orogenic belts seems to be linearly related to the elevation difference between the summits and the forelands. The present day orogenic plateaus are 3.5 to 5 km higher than their forelands, independently of their width. This suggests that the height of the collisional plateaus represents a maximum possible. If the orogenic belts are in isostatic equilibrium, this maximum provides a first order proxy for the maximum thickness the continental crust can reach. Calculations are about double thickness.

Erosion

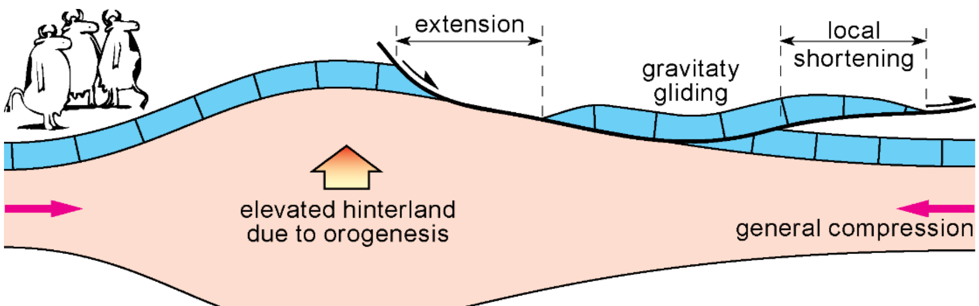
The mountain belt becomes also a region of erosion, which produces **denudation** and **exhumation** of deep levels and supplies sedimentary basins. On one hand, the removal or addition of material at the surface affect the load and alter subsurface stresses that drive the deformation. On the other hand, the rates of erosion and deposition depend on relief and other tectonic features. Since erosion tends to remove weight from **uplifting** blocks and deposition tends to add weight to subsiding blocks, surface material transport and tectonic deformation are often in a positive feedback relationship.



Long-term evolution of a mountain system

Gravity gliding

Gravity slide tectonics embraces phenomena whereby large and relatively coherent blocks or slabs, bounded below by distinct zones or planes of detachment, are translated laterally under their own weight. The process typically occurs in the upper few kilometres of tectonically active areas where differential uplift and subsidence provide the gravity potential to permit down-slops flowage of the rock sheet away from the raised orogenic interior. Thrusts develop at the toes of the sliding sheet.



Geometry of gravity slide tectonics

Gravity spreading involves deformation of the entire rock mass, the elevated hinterland collapsing down and forcing its leading edge away, towards and over the foreland. This concept is inherent to orogenic collapse.
A gravity-driven system has three parts:

1. The **breakaway** normal fault, at its back.
2. The detachment zone, often along a single bedding plane.
3. The thrust ramp and transgressive fault that steps up to foreland surface.

Collapse of orogenic systems

Collisional mountain belts are formed by tectonic compressional forces. When these horizontal forces cease acting, the elevation of the mountain or plateau will relax by one or both of two processes:

- (1) erosion which removes the near surface rocks and allows uplift and
- (2) gravitational collapse of elevated topography.

In effect, many orogenic belts terminated their compressional evolution with throughout extension. Horizontal extension is possible where the largest principal stress is subvertical, i.e. where the lateral collisional forces are suppressed by the vertical body forces. Such as state of stress occurs in high topography areas. There, the weight of the mountains exceeds the yield strength of the buried, heated and weakened continental crustal root. Besides, the lithostatic pressure in the crust beneath thick mountains is higher than at equivalent depths beneath adjacent lowlands. Because of this pressure gradient, highlands become gravitationally unstable. Their weight is the body force that tends to push the crustal rocks towards the lowlands, which causes crustal thinning and associated loss of elevation. **Collapse** accordingly occurs as a late result of crustal thickening and excess topography and can lead to the total loss of elevated topography without erosion.
The geometry and kinematics of extensional collapse depends on many parameters. It is in general directed away from the elevated hinterland towards the low forelands. Extension is partitioned into ductile flow in the hot, ductile lower crustal levels and listric detachment faulting in the colder, brittle upper crust. The process helps exhuming deep structures of the mountain belt in dome-shaped elevations that provide windows into rocks that were deeply buried during collision: the **metamorphic core complexes**, which form horst-ranges separated by **intermontane basins.** Thinning of the upper crust causes decompression, hence partial melting of the deep crust. Consequently, migmatites and granitic magmatism are characteristic features of collapsed orogens.
Where the elevation is compensated by a crustal root, extension is confined to the crust itself. Where the elevation is compensated by a low-density subcrustal body, collapse may be triggered by the delamination of the lithospheric mountain root. This lithospheric root, the **keel**, will be replaced by

hot asthenospheric material, which may cause a late orogenic high-temperature metamorphism, crustal melting and granitic magmatism. Extension then affects the entire lithosphere.

Collisional mountains through geological times

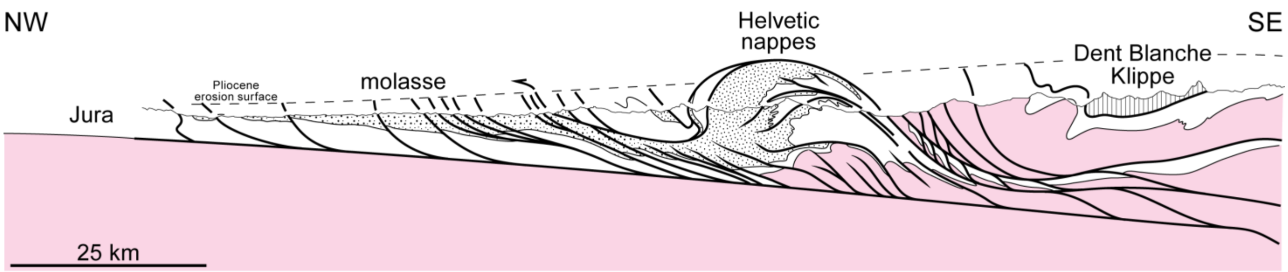
A corollary of the collapse argument is that lithospheric strength is an important factor limiting the mountain height. Since temperature is a major parameter controlling the rheology of rocks, the temperature history should affect the orogenic history of the Earth. Indeed, in addition to the primordial heat (the internal heat energy accumulated in the planet during its accretion, differentiation and core formation), radioactive decay contributed about three times as much heat at the beginning of the Archean and almost twice as much in the Early Proterozoic as it produces today. This evolution of the heat budget was used to argue that there was no plate tectonics on Earth before ca 3.2 Ga, when heat transport and cooling were dominated by advection via profuse volcanism on a vigorously convecting mantle. When did continental lithosphere began to form is still open question, but heat loss by conduction and radiogenic heat made young continents warmer, hence weaker than they are today. Therefore, mountain ranges of the time collapsed when their elevation (weight) was lower than elevations reached in modern collisional systems. Enhanced mantle convection possibly favoured delamination. Weaker continents also means a different style of shortening deformation, which is another contentious topic.

Balanced cross sections

A fundamental preoccupation in upper-crustal, thin-skinned thrust systems is how folds are related to the fault geometry. A complete understanding of map-scale structures involves the construction of cross sections, preferably oriented perpendicular to the regional strike. **Balancing** these cross sections provides geometrically constrained structural models and weed out geometrically unworkable interpretations.

Concept

The basic, strongly consequential rule assumes conservation of volume in three-dimensions, which reduces to conservation of area in two-dimensions. Solutions to cross-sections are tested by **retro-deformation**. This test means that if all shortening represented by fault displacements and folds in the section is removed, the layers should restore to a pre-deformation configuration with no large gaps or overlaps in strata. If the rock volume remained constant through the deformation history, the pre-deformation, **restored** section shows the original stratigraphy with the fault trajectories. With this test, a **viable,** balanced cross-section is actually a pair of cross-sections: one showing where rock units are now, and one showing the configuration of these units before deformation.



Thin-skin tectonics interpretation - balanced profile through the Alps
after Boyer & Elliott 1982 *Am. Ass. Petrol. Geol. Bull.* **66(9)**, 1196-1230

There are two main reasons for restoring cross-sections. First, restoration with geometric models helps to evaluate whether the section is structurally reasonable or not; if assumptions are correct, it must be geometrically possible to undeform the section to its initially undeformed state. Second, the section restoration provides a kinematic model of the progressive development of the studied fault system. The kinematic path is the third, important element of section balancing, linking in a credible manner the pre- and post-deformation sections.

A geologically **admissible** cross-section respects the structural style of the region, which means that it depicts structures that look like those observed in the field, and eventually in seismic profiles. A cross-section that respects both the style and the restoration constraints is a **balanced** cross-section because length and thickness of rock units are equal in pre- and post- deformation profiles.
In reality, this is only possible in sections across sedimentary sequences that have not undergone internal strain and, therefore, this exercise is limited to the margins of major orogenic belts, the foreland fold-and-thrust systems.

> Attention: A balanced cross-section is not a unique and correct solution. An unbalanced section can be correct if there is out-of-plane, lateral strain, there are volume changes, or thin-skinned tectonics simply does not apply. However trying to balance sections enables avoiding gross structural errors.

Assumptions

Balanced cross-sections are generally constructed for deformed sedimentary sequences where the stratigraphy is known. In metamorphic and polyphase terranes the construction is practically impossible.

Mechanical stratigraphy

Intuitively and empirically (with the hammer), sedimentary layers such as shale are known to be weaker than limestone and sandstone layers, for example. This strength difference is even more contrasted if high fluid pressures further weaken the weak layer. Under low temperature, brittle ("Mohr-Coulomb") behaviour, the angle between a fault plane and the maximum principal stress is smaller for weak materials than for stronger materials (lecture rheology, strength profiles of the lithosphere). Therefore, a sedimentary sequence will fracture along angles that depend on the shear strength of the rock. This theoretical summary concludes that stratigraphy plays an important mechanical role, with fault planes cutting a strong rock at a higher **cut-off angle** to bedding than in weak rock. A master fault plane across a sedimentary sequence thus links segments that step-up across strong layers and usually longer segments parallel or subparallel to bedding in weak layers. The result is a network of ramps and flats separating the hanging wall from the footwall. The hanging wall slides along the flats and moves up the ramps.

Transport direction – plane strain

The cross section is parallel to the transport direction. All motion has been parallel to the plane of the section and there is no volume change. Volume conservation is respected by folding, imbrication and décollement of the deformed sequence (thin skin tectonics), or folding, imbrication and basement faulting (thick skin tectonics).

Conservation of lengths / areas

The fundamental assumption of plane strain deformation precludes movement of material in or out of the cross-section; consequently, there are no holes in the section before and after deformation. The section is balanced between two lines. One **fixed**, reference **pin line** is chosen to pass perpendicular to bedding (generally vertically) through the undeformed, autochthonous footwall section of the most complete stratigraphic sequence (the **template**), including the basement. This line is usually placed in the undeformed foreland; it serves as a marker from which bed lengths are measured. The other, the **loose line**, is placed perpendicular to bedding in hanging wall thrust sheets. These two lines delimit the area that remains the same in the deformed and restored sections. In general, loose-lines should stay straight. If not, the tilt or offset (sense of shear) should be consistent across the displaced hanging wall.

Kink-style passive folds

Since there is no internal strain in the layers, folding is assumed to be parallel. It is then convenient to construct sections using a kink-fold geometry, i.e. folds with angular hinges, straight limbs of constant thickness on either side of the hinge, and axial planes bisecting the inter-limb angle. Large, smoothly curved fold may be approximated by several neighbouring kink-folds.

Matching cut-off points

Because a thrust has displaced hanging-wall rocks from correlative strata in the footwall, there is a unique hanging-wall cut-off and a corresponding footwall cut-off. In a balanced cross-section, hanging-wall and footwall cut-offs must restore points that formerly were immediately adjacent to one another across the fault. Note that at these points the cut-off angle remains constant and is measured relative to bedding, not to the horizontal.

Structural compatibility

For each hanging-wall ramp there must be a corresponding footwall ramp. Similarly, for each hanging-wall flat there must be a corresponding footwall flat. This is known as the template constraint. There are no cavities or holes in the section before or after deformation.

Unstrained footwall

The method also assumes that the footwall remains undeformed and that there is no ductile deformation during thrusting.

Methods

There are two basic methods of drawing balanced cross-sections. Both techniques are used as a mutual crosscheck on the admissibility of the section. Practically, the interpretation progresses by tentative solutions (trial and error).

- The *constant line length restoration* assumes that thrust-related folds are produced by a flexural-slip mechanism so that layer thickness does not change. If layer thicknesses remain constant during deformation, it is only necessary to maintain the bed length to keep the same area, as volume equivalent in the present and restored states. This impels that two layers at different depths have the same length. This approach is particularly appropriate for parallel folding of competent sedimentary layers.
- The *constant-area balancing restoration* is used in areas where deformation has changed bed length and thickness, such as the fold-and-thrust systems that involve similar style folds and cleaved rocks. This approach is appropriate for incompetent sedimentary layers. The total area of stratigraphic intervals in a cross section is the same in the present and the restored states. This method is more tedious but less constraining than line-length balancing.

Construction

- Working on sections with the same vertical and horizontal scales is of primary importance to prevent inconsistent geometrical interpretations.
- Gather all accessible data (geological observation, boreholes, seismic profiles, etc.) to determine the stratigraphic thickness and construct a stratigraphic template (a restored, pre-deformation profile).
- Construct the topographic profile of a section perpendicular to fold axes inferred to be orthogonal to the tectonic transport direction (i.e. section parallel to the transport direction, along which down-plunge projections are safe information).
- Assemble along the section line the lithological boundaries, faults, strikes and dips from geological maps, and incorporate well-log and projected data.
- Define the pin lines where there is no inter-bed slip (beds are stuck together). Usually, these points are located in flat foreland or on the axial plane of folds.
- Construct a first, trial profile and restore it step by step by plotting the measured bed lengths and thickness of deformed strata from the fixed pin line (the foreland autochthonous section) towards the hinterland, i.e. towards older and older ramps and related folds.
- Extrapolate structures at depth, producing a restored section at the same time as constructing the balanced section. The restored section, with all the pieces back in their places, should have no

gaps or overlaps and the staircase fault trajectories should be plausible, with ramps originally dipping 30° or less and flats lying in the right layers.

- Use fold shapes to infer fault position and orientation at depth following the predictable geometric relations between folds and thrust trajectory:
  - i) Define panels of constant dip and project all contacts into the air as much as into the subsurface.
  - ii) Determine the orientation of axial surfaces of folds by constructing the bisectors of adjacent dip panels.
  - iii) Project folds to depth or into the air where eroded. Where two axial traces merge downward, the result is a single hinge that bisects the chevron fold. Construct a new axial surface that bisects the limbs of the chevron fold.
  - iv) Determine whether fault-bend folds or fault-propagation folds are present.
- Select the solutions that best honour the geological data after repeated adjustments on section.
- Check validity of the retro-deformed state. Any remaining imbalance should be explained (e.g. salt dissolution, non-plane-strain deformation, etc…).

Note

Deformed (oblique to bedding) loose lines are often obtained in constant length restoration. They point to restoration or deformation problems (bedding parallel shear, flexural slip….) that require step by step hand correction (needs some inspiration and courage to give up after several failed attempts to balance and rebalance). Step by step means from pin line backward, from ramp to ramp, to minimize error propagation.

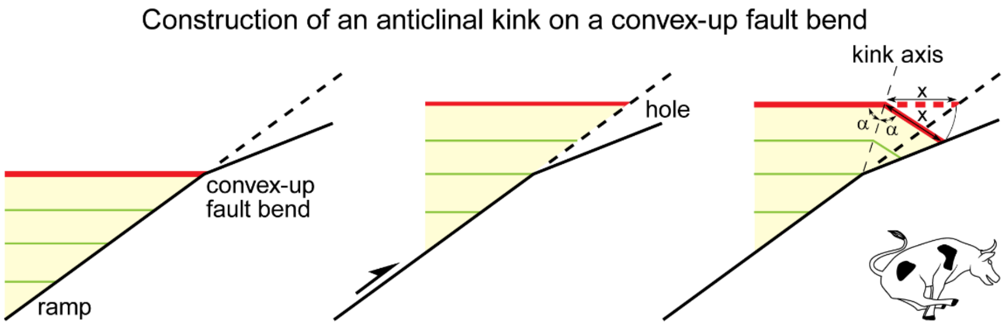
- *Construction of fault-bend folds*

Common kinematic models and balancing techniques applied to passive folds make the assumption that ramps and flats are angular and use the **kink construction**. The kink construction further assumes that fold hinges are angular between limbs of uniform dip and constant thickness of strata (parallel folds). This is acceptably justified in shortening systems because it supposes flexural slip (bed-on-bed movement), which is generally uncommon in extension systems. In addition, geologists have commonly observed that even where folds appear as broad curves, the curves usually consist of many small, straight-line segments. The construction involves several working steps:
- Location of regions of homogeneous dip (**panels**).
- Location of axial surfaces between these panels.
- Location of fault discontinuities where axial surfaces terminate.

The principle of fault-bend folds is that horizontal motion along flats causes the layers to passively ride up ramps and roll through axial surfaces, two of which are affixed to the top and bottom flat-ramp connections.

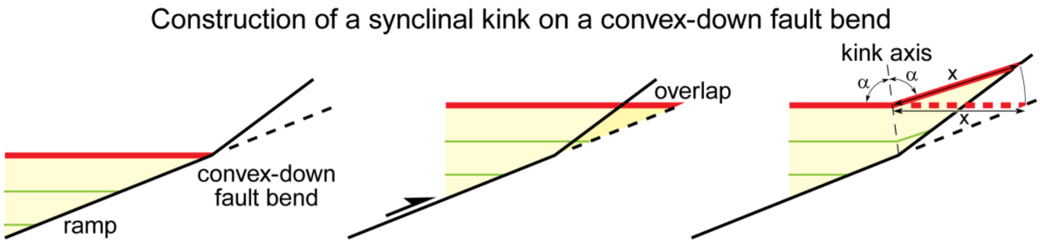
Convex-up fault bends (top of ramp)

Upward movement of the hanging-wall creates a hole that deformation of the hanging-wall must fill. The axial plane of the resulting fault-(kink) bend is placed such as it bisects the antiformal fold (kink construction constraints) and that lengths are preserved in the collapsed limb (line-length balancing constraint).



Convex-down fault bends (ramp bottom)

Upward movement of the hanging-wall creates a ramp-top hole that antiformal rotation of the hanging-wall must fill. The axial plane of the resulting fault-(kink) bend is fixed to the ramp-to-flat bend and is inclined such as it bisects the antiformal fold (kink construction constraints) while bed lengths are preserved in the collapsed forelimb (line-length balancing constraint).



To respect such geometrical constraints, the beds passing through the top-of-ramp axial plane experience bedding plane slip, hence some shear in the frontal limb which consequently dips steeper than the ramp (the cut-off angle of the front is larger than the initial cut-off angle against the ramp). The exact solution for these angles is cumbersome trigonometry and can be found in graphical solutions developed in Suppe (1983) *American Journal of Science* **283(7)**, pp. 684–721, Fig. 7 p 694. Since the front limb undergoes some internal deformation, the exact solution implies that slip on the upper flat is less than on the lower flat. Alternatively, the construction is a nice approximation but the nature does not follow mathematically strict solutions.

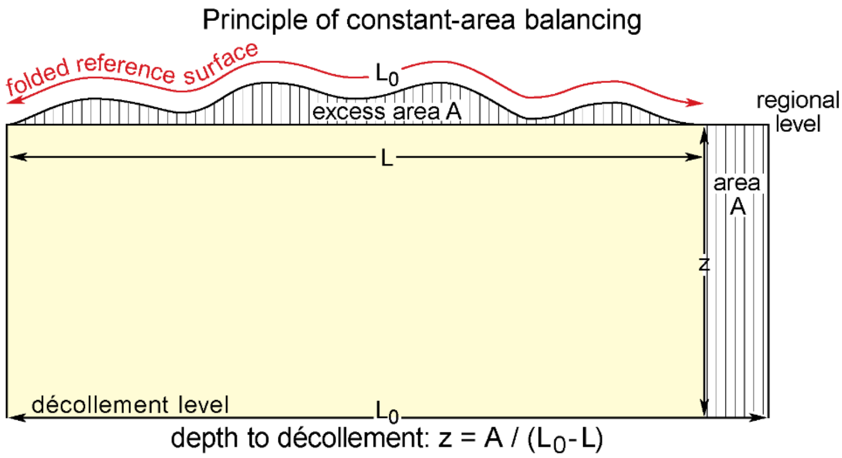
Convex-down fault bends (ramp bottom)

From flat to ramp movement of the hanging-wall forces synformal fault-(kink) rotation of the layers through an axial plane fixed to and bisecting the flat-to-ramp angle. The same kink construction and line-length balancing constraints as for convex-up fault-bends applies.

Detachment fold

Because detachment folds are filled by much thickened, incompetent rocks whose length and thickness are not conserved, the kink construction tends to produce unrealistic-looking sections of such folds. Area balancing is required. The arc length L_0 and the pin length L of a reference bed are measured. The **excess area** A uplifted above the pinned, undeformed height of the bed should be equal to the area of that bed reduced by shortening displacement along the décollement level. Knowing the horizontal shortening (the difference with the initial length L_0 obtained by straightening out the arc length of the fold), it is possible to calculate the depth z to the décollement level.

$$A = z\left(L_0 - L\right)$$

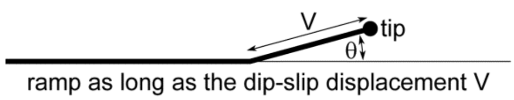


Fault-propagation folds

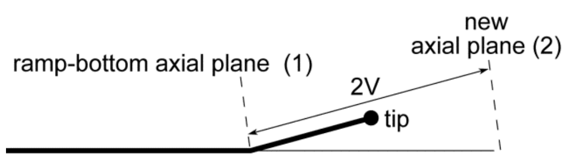
Construction of fault-propagation folds combines area-balancing techniques as for detachment folds (in that the fault displacement decreases to zero in the direction of transport) and line-length balancing as for fault-bend folds (because folding takes place on a ramp). In the latter case, the amount of shortening by folding in strata higher than the fault tip must be exactly the same as the amount of shortening by thrusting at the basal detachment.

Trigonometric solutions for balancing idealized fault-propagation folds by kink construction have shown that the slip on the fault is exactly half the length of the backlimb of the folds. To sketch a fault-propagation fold using the kink construction:

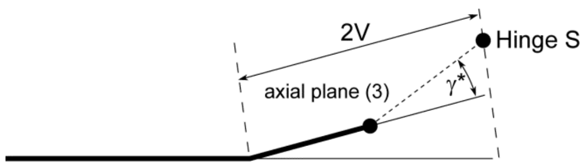
- First draw a flat and a ramp having a length equal to the amount of slip V, to the tip point T.



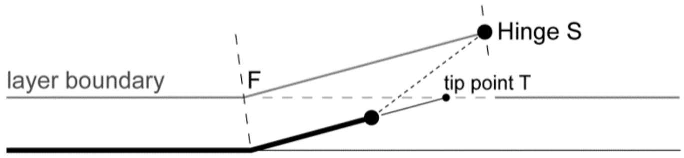
- Bisect the angle at the base of the ramp to sketch a fold axial plane. Sketch another fold axial plane, ahead of the first by a distance 2V.



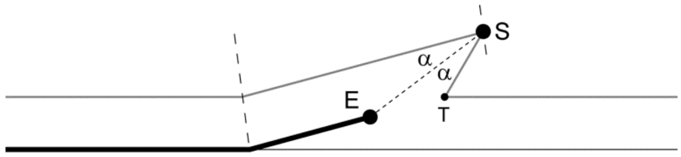
- A third fold axial plane extends from T, more steeply than the ramp (making an angle γ^* to the ramp). Where it intersects the second axial surface, mark a hinge point S. To make the construction work exactly, the angle γ^* has to be taken from the graph generated by Suppe and Medwedeff (1990) *Eclogae geologicae Helvetiae* **83(3)**, pp. 409–454.



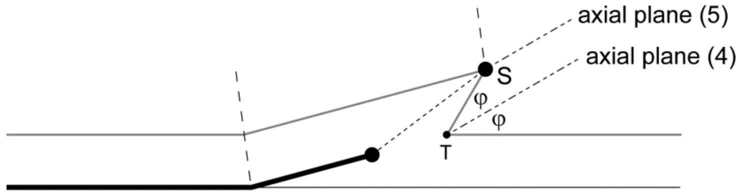
- Construct the lowest folded layer boundary unaffected by faulting, which passes through S.



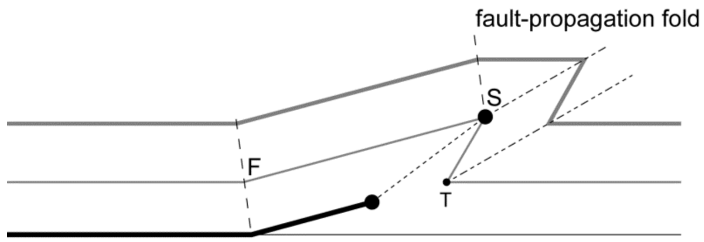
- Extend the ramp forward to a tip point T that lies on the same level as F. The fault slip declines linearly to zero up to the old tip E.



- Construct the forelimb between S and T. If the value of γ* is precise, then axial surface 3 should bisect the angle of the completed antiform.



- Draw a fourth axial surface bisecting the angle between the forelimb and the horizontal strata ahead of the fold. A fifth axial surface extends forward from S at the same angle. Construct higher strata with a horizontal upper limb between axial surfaces 3 and 5.
- For the fold to balance correctly, all the axial surfaces should bisect the folds, and the shortening by folding (B+C-A) in the top of the section should exactly balance shortening by thrusting (S) at the bottom.



- For the fold to balance correctly, all the axial surfaces should bisect the folds, and the shortening by folding (B+C-A) in the top of the section should exactly balance shortening by thrusting (S) at the bottom.

Although geometrically exact, this model has been questioned on kinematic grounds. For a fault-propagation fold to maintain this geometry while the fault propagates, it is necessary for strata to distort and undistort as it continuously rolls through the fold axial plane. Examination of real folds seldom shows evidence that the rocks have been through this type of strain history in a rolling hinge. Hence alternative geometries have been proposed, in which axial planes remain fixed for periods of time. For these to work, it is necessary to relax the strict constraints of the kink construction, and to use area balancing where layers change thickness.

Effect of footwall deformation

Footwall deformation strongly modifies the passive footwall model in which the hanging wall is kinematically active and elevated. If hanging wall and footwall rocks have similar rheologies, and if the ground surface is high above the thrust plane, then the overall shortening should affect both sides of the thrust. A folded footwall changes the general displacement field in both the horizontal and vertical directions.

Exercise

Draw and study the development of a fault-bend fold involving folding of the footwall, symmetric to the fold bend with respect to the thrust plane.

Analytical consideration

The simplest analysis assumes that all horses have the same size and shape. Again, excluding any ductile deformation:
The total slip recorded by a duplex is:

$$L_0 - L = n(\ell - s_n)$$

Where L and L_0 are post-thrusting and initial length of a marker bed, respectively, n the number of horses, ℓ the length of bed within individual horses and s_n the linear dimension of a horse measured parallel to the floor thrust.
Shortening is thus:

$$\frac{L - L_0}{L_0} = \frac{n(s_n - \ell)}{n\ell} = \frac{s_n - \ell}{\ell}$$

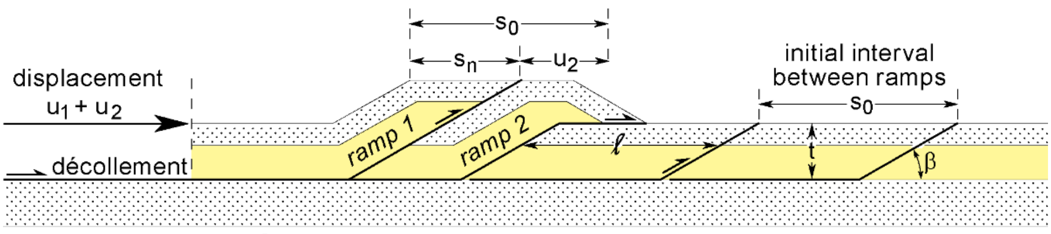
This equation shows that total slip increases with an increase in the number of horses n, but shortening is independent of n. Taking β the angle between the floor thrust and a ramp, simple geometrical constructions show that

$$t/s_n = \sin\beta$$

with t the thickness of beds included in a horse. Shortening becomes:

$$\frac{L - L_0}{L_0} = \frac{t}{\ell \sin\beta} - 1$$

which shows that for the same values of t and l, the absolute value of shortening varies with β. The limiting value is obtained when shortening is 0. For example, if $t/\ell = 1/4$, β cannot be smaller than 14.5°.

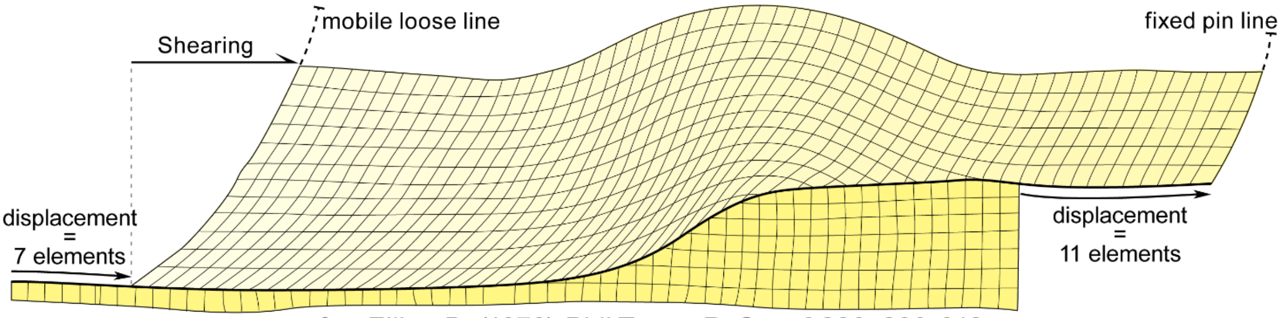


Geometrical variables of an imbricate zone

Duplexes exhibit a variety of forms, depending on the amount of displacement of the individual horses. If the initial spacing between the imbricate faults measured parallel to the floor thrust is s_0 and the displacement on each imbricate fault is u_n, then for
$u_n < s_0$ Imbricate faults dip towards the hinterland in hinterland dipping duplexes.
$u_n \approx s_0$ the duplex forms an antiformal stack
$u_n > s_0$ With large displacements, horses can pass over and beyond underlying horses and give rise to foreland dipping duplexes.

A technical tip

Use scissors, cut your section along every fault plane and every axial plane of the kink folds. Undeform by placing all pieces (panels) in the right order and position on your desk and tape them down together. Due to flexural slip or other strain processes, there always be some small gaps at kinks. Minimize them.



after Elliott D. (1976) *Phil.Trans. R. Soc.* **A283**, 289-312

Caution warning

Balanced cross-sections are a model that relies on some issues:
- Availability of sufficient field, well log and geophysical data to ensure accuracy of cross-section models.
- Strike-slip motion across the section line makes it difficult to produce viable sections.
- The method requires the existence of a relatively flat décollement fault and therefore produces such faults.
- The model may infer the existence of faults that, lacking surface deformation, may be difficult to evaluate.
- There are always several solutions that balance sections. Balancing section is only a convenient approximation to reality.

Features which interact with thrusting

Older basement features, facies and thickness changes in the sedimentary succession.

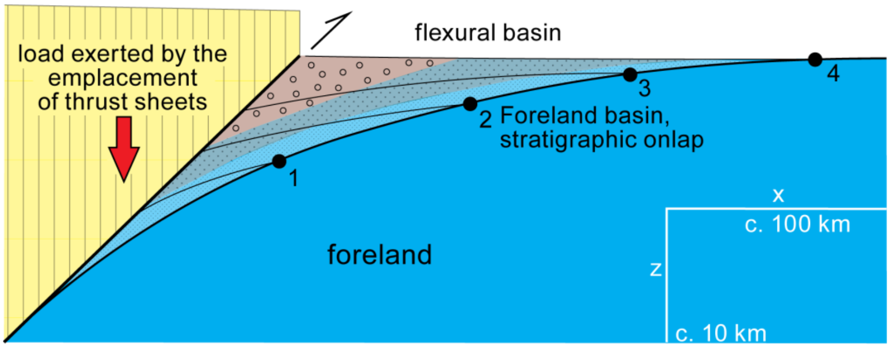
Sedimentation in thrust-zones

Sedimentary basins can form in a compressional context. Sedimentary wedges are typical for regions with syntectonic sedimentation.

Foreland basin

A **foreland basin** forms in front of an orogenic belt because the orogenic mountains, which are transported on a major, emergent thrust and gain weight as they thicken, load the footwall lithosphere. Consequently, the footwall lithosphere bends down into an asymmetric depression adjacent and parallel to the mountain belt. A relatively low-elevation bulge delineates the external border of foreland basins. Material eroded from the high topography mountain is deposited in this syn-orogenic depression, which deepens towards the mountain belt (**flexural isostasy**).

Migration of the foreland basin towards the foreland



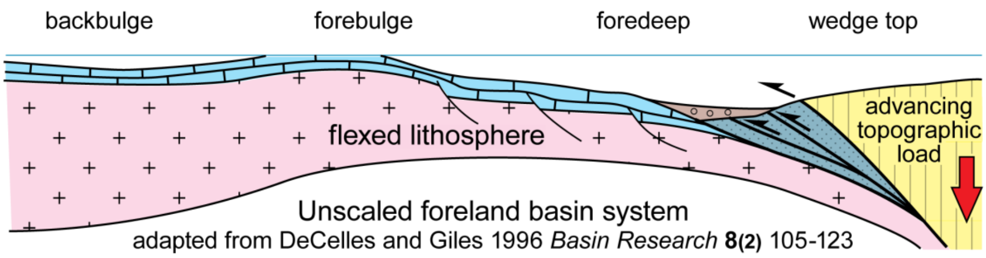
There are two types of setting:
Pro-foreland basins formed on the subducted / underthrust plate.
Retro-foreland basins formed on the hanging wall plate.
Foreland basins of both settings are wedge-shaped, with a convex-up floor, thick along the mountain margin, thinning to zero on the external side against the peripheral bulge. The foreland basin consists of four depositional zones that move with time towards the orogen:

The back-bulge covers the wide zone of relative low-land behind the bulge, away from the orogenic belt. It is the site of platform sedimentation.

The forebulge where minor flexural surface uplift takes place.

The foredeep, ahead of the thrust belt, where progressively coarser sediments record progressively accelerating subsidence.

The wedge-top on the frontal part of the orogenic wedge, where foreland sediments may become incorporated into the thrust belt.



Unscaled foreland basin system
adapted from DeCelles and Giles 1996 *Basin Research* **8(2)** 105-123

Simple mechanical models of the lithosphere reproduce well the form and history of foreland basins. The deflection at any particular point varies as a function given by:

$$z = z_{max} e^{-x/\alpha} \cos(x/\alpha)$$

where x is the horizontal distance and z is the deflection.
The quantity α is given by:

$$\alpha = \sqrt[4]{(4D/g\Delta\rho)}$$

Where D is the flexural rigidity of the lithosphere (usually about 10^{24} Nm) and $\Delta\rho$ is the density contrast between the asthenosphere below and whatever is on top of the plate (e.g. water).
Footwall flexure may be accentuated by the weight of sediments, which are the erosion product from the growing mountain chain.

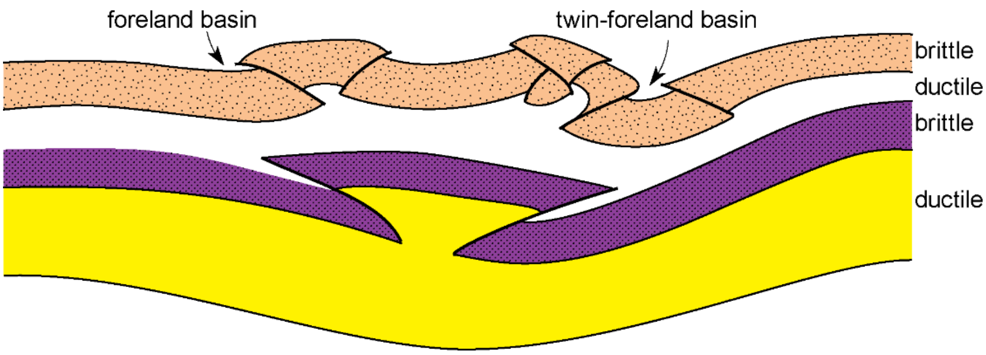
Sediment history

While the footwall plate moves towards the mountain front, any given point experiences successively platform sedimentation, minor uplift and possible erosion at the peripheral bulge, progressively accelerating subsidence and deposition of gradually coarser sediments in the foreland basin and eventual integration into the fold-and-thrust belt. If subsidence is faster than sedimentation, the basin is **underfilled** and generally hosts deep-water sediments (typically turbidites, the so-called flysch in the Swiss Alps). If sedimentation balances or even overrates subsidence, the foreland basin is **overfilled** with shallow marine to fluvial clastics.

Twin-foreland basin

In contrast to foreland basins, which form by footwall flexure under a single thrust twin-foreland basins (classically called ramp valleys or full ramp basins) form by motion of paired reverse faults toward each other.

Development of compressive basins
in four-layer sand-silicone analogue experiments
after Martinod (1991) *Mém. Doc. Centre Arm. Et. Struct. Socles* **44**, 283 S.

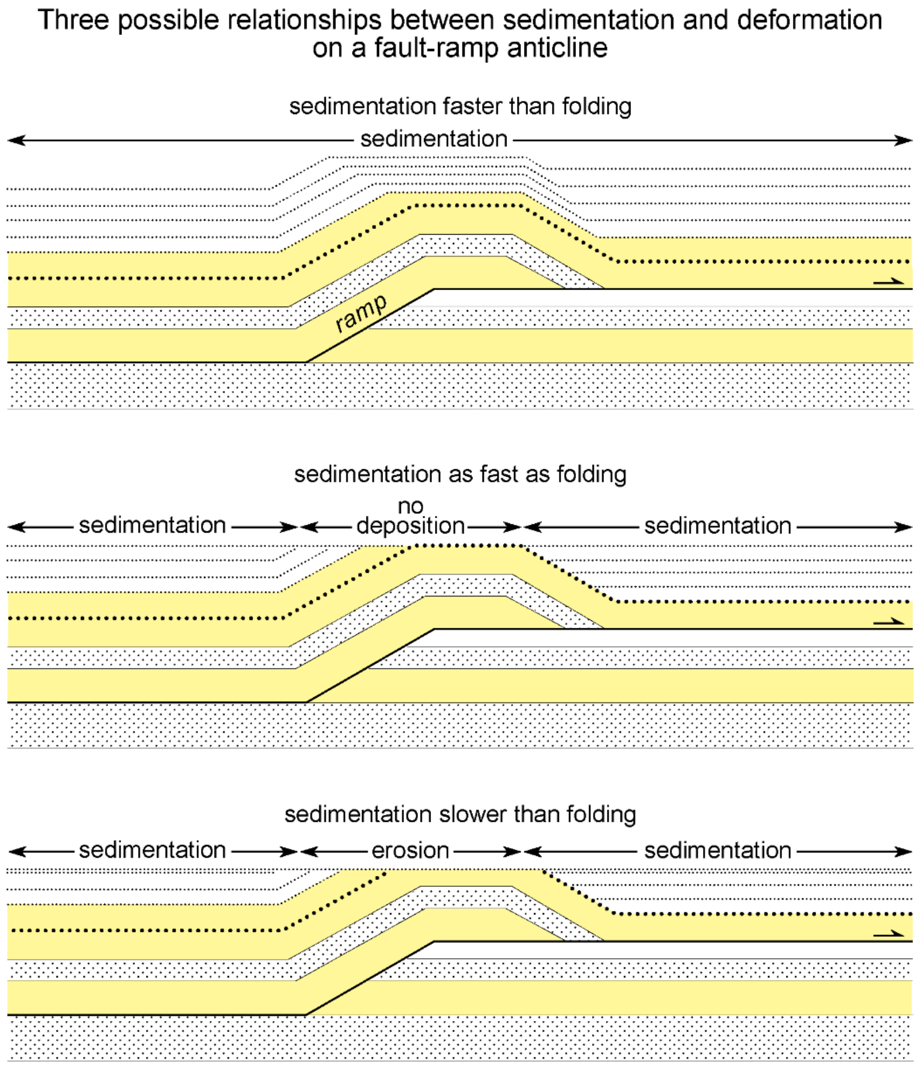


If these conjugate faults are closely spaced, flexure is inhibited and the basin in between has a flat bottom. With increasing shortening, the twin-foreland basin is pushed down, a process that is accelerated by sediment load. After a large amount of shortening, the bounding thrusts may meet and thus totally hide the basin from surface exposure.

Syn-sedimentary thrusts

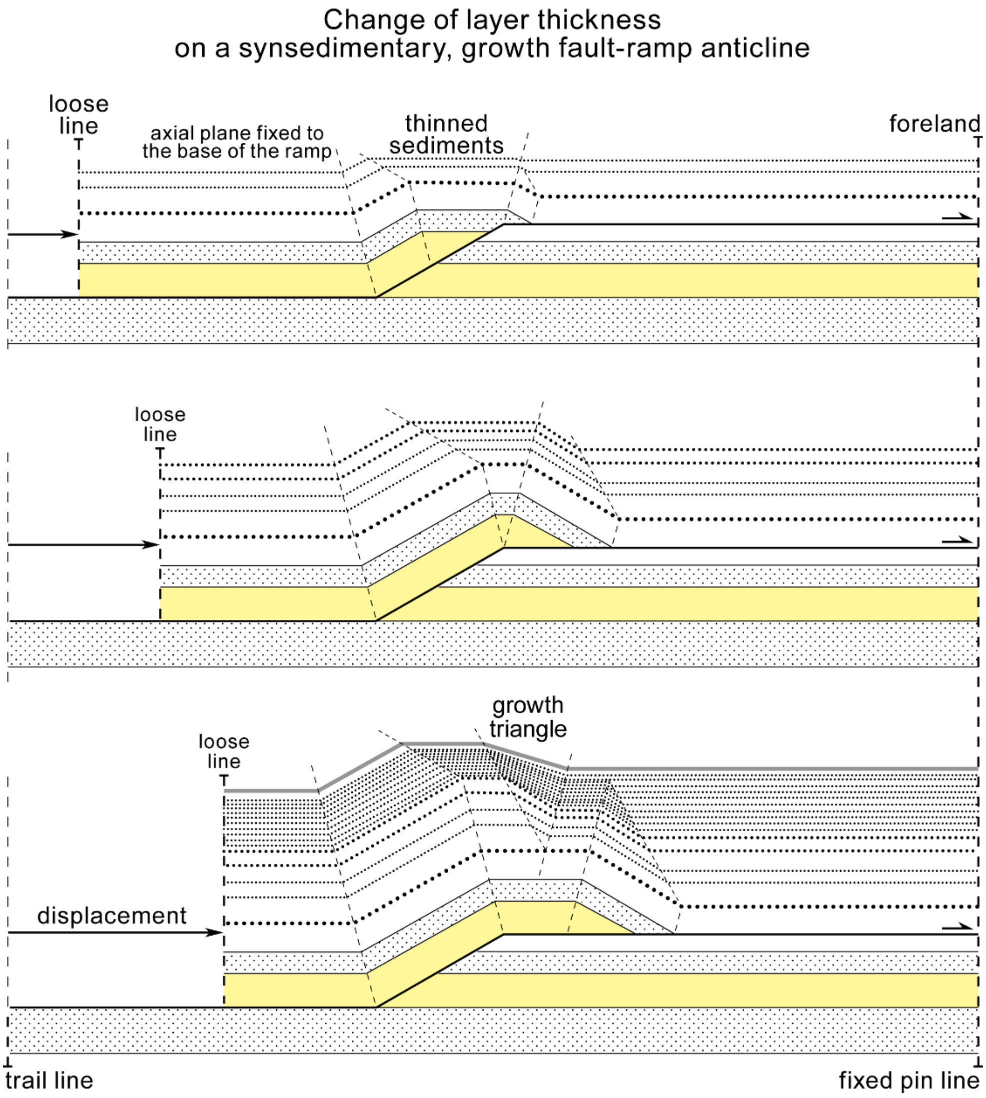
Analogue modelling suggests that syntectonic sedimentation controls the number and the dip of faults in the upper part of thrust systems, i.e. synkinematic sedimentation modifies the thrust wavelength and the major propagation of the deformation. Single low-angle thrusts develop for low sedimentation rates; series of fanning and steeply dipping thrust are obtained for high sedimentation rates.

Syn-sedimentary thrusts are particularly common and strongly control the sedimentation pattern on submarine foreland basins and accretionary wedges.



Growth folds

Synsedimentary folds (growth folds) may form above thrusts propagating over a ramp. Differential sedimentation between the raising, high crest of the fault-propagation anticline and the adjacent, relatively subsiding and low synforms results in beds (**growth strata**) wedging across the fold limbs. The wedge shape is primarily controlled by the relative rates of sedimentation and fold amplification.



Inversion Tectonics

Basin inversion occurs when pre-existing graben-bounding and within-basin normal faults become reactivated in compression as reverse faults. Inverted basins are common in orogenic forelands where compression due to plate convergence and gravity push of topography acts on older basins (e.g. Carboniferous-Permian troughs in the Alpine foreland).

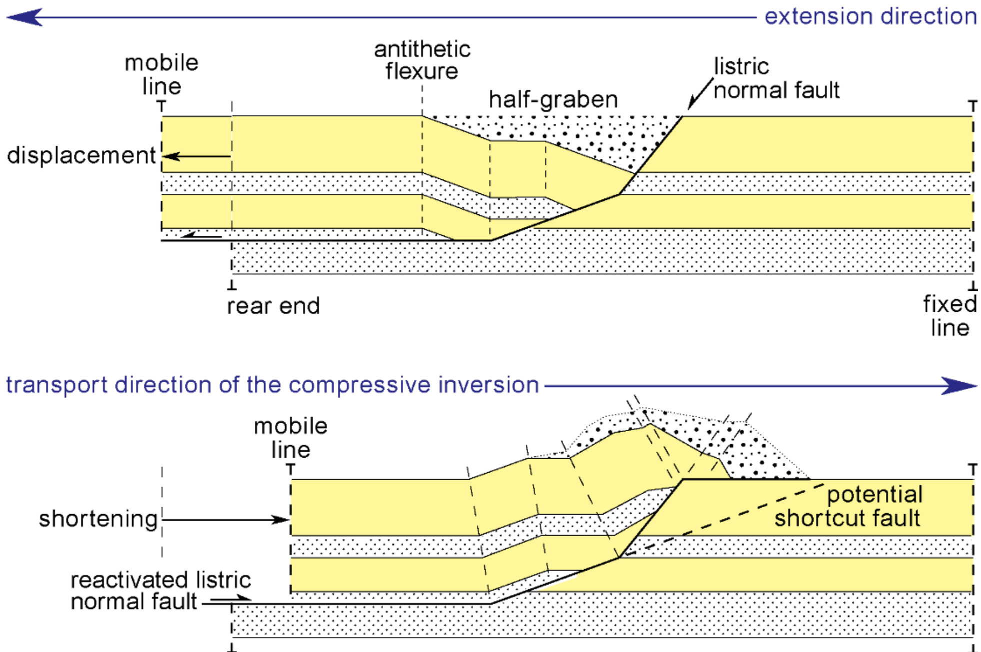
Three main models of inversion have been postulated from seismic, field and analogue studies:

(1) Fault-reactivation model. Basin inversion takes place by the progressive, top-to-bottom reverse reactivation of normal faults, with formation of derivative splay and shortcut faults, so that the changeover, zero-movement point between reverse above to normal below, moves down the fault surface. Fault-reactivation is generally selective, with the shallowest dipping faults reactivating preferentially unless fluid pressure reduces cohesion on steeper faults.

(2) Thin-skinned model. A steep fault (dip > 60°) is difficult to reactivate unless it has a very low cohesion and sliding friction, as under very high pore fluid pressure. Therefore, normal faults, because of orientation, buttressing or other reasons, undergo only minor reactivation or are not reactivated. Instead, it may be energetically more favourable to deform intact rock in the rigid footwall block. Commonly, a new **footwall shortcut fault** in the rigid footwall block dips at a lower angle than the older fault and shortening of the basin-fill is accomplished by formation of new thin-skinned thrusts that cut older structures.

(3) Buttress model. Basement rocks in the footwalls of steep normal faults form mechanical barriers to the reverse reactivation of normal faults and to thin-skinned thrusting along or near the basement-cover interface. As a result, shortening of basin-fill by folding and back-thrusting is concentrated in

the hanging walls of the normal faults, and may be accompanied by the minor reverse reactivation of the upper parts of those faults.



Basin inversion

If the top of the pre-rift sequence returns to its pre-rift level during inversion, the basin has undergone **total inversion** and syn- and post-rift sediments are uplifted above their original depositional level. Inversion involves a greater component of vertical displacement than generally occurs in thrust tectonics. Complexity is added to these models when the directions of extension and compression are not parallel. In such cases, inversion involves a component of strike-slip movement leading to oblique or transpressional closing of extensional basins. Oblique deformation may be taken up by the formation of newly formed inversion structures that are not coaxial with extensional structures, or by the strike-slip or oblique-slip reactivation of extensional faults. The resulting fault geometry may be difficult to distinguish from **flower structures**. In a three-dimensional strain field, these oblique faults act to partition inversion into compartments with different structural styles. Basin inversion may be **positive** or **negative**. Reactivation of basement faults may also produce segmentation of inversion structures.

The degree and timing of inversion can be quantified by using **displacement-distance plots** and the **inversion ratio**.

Conclusion

Thrust faults, which are often accompanied by back-thrusts, bound thrust sheets. Thrust belts are the result of horizontal compression at convergent continental margins. Where two continents collide, burial and deformation of the continental crusts produce mountain belts, which are large thrust systems. Deformation may involve basement (thick-skinned), or be limited to the sedimentary cover, which is detached from the basement (thin-skinned). Fold-and-thrust belts are typical of most mountain belts. They reflect shortening of the upper crust. Thrust sheets deform internally by folding: two mechanisms can be recognised - fault-propagation folding and fault-bend folding. Ramp anticlines are fault-bend folds in the hanging-walls of thrust faults, formed when thrust sheets are carried over flats and up ramps. Thrust faults form sequentially with regular thrust sheet lengths and spacing, in normal-sequence from hinterland to foreland and characteristically form duplexes. Complications such as syn-tectonic sedimentation or variations in basal friction cause out-of-sequence thrusting. The arrangement of flats and ramps and their connections like in duplex structures depends on the mechanical stratigraphy. Thrust faults may be separated by folds (soft-links) or by oblique or lateral ramps (hard-links), and when the separation becomes very large, by tear faults. Seismic imaging is commonly a problem due to the structural complexity and mountainous terrains.

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