CHAPTER EIGHTEEN

Fold-Thrust Belts

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18.1 INTRODUCTION

Picture yourself hiking among the glaciated spires of the Canadian Rockies (Figure 18.1). Where did the beds of sedimentary rock forming these mountains come from, and how did they end up exposed on cliffs 2 km above sea level? The sediment composing the beds originally accumulated on the floor of a sea, tens of kilometers to the *west* of their present location. Eventually, the sediment was deeply buried until it lay several kilometers *below* the Earth's surface. Thus, these strata had to move large distances both horizontally and vertically to get to their present location in the Canadian Rockies. After uplift, erosion by rivers and glaciers carved the rugged cliffs on which the strata crop out today.

Formation of the Canadian Rockies involved displacement on **thrust faults,**[1] dip-slip faults on which hanging-wall blocks slide up the fault surface. Geologists refer to the bodies of rock that move during thrusting as **thrust sheets** or **thrust slices.** In the case of the Canadian Rockies, the stress driving thrust-sheet movement was probably generated by convergence and/or microplate collision along North America's western border during the Mesozoic and Early Cenozoic Sevier and Laramide Orogenies. As a consequence of thrusting, once-horizontal beds of sediment may become tilted or folded and may develop tectonic

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[1] We'll introduce other names for such faults later in this chapter—unfortunately, fold-thrust belt jargon has become quite complex!

FIGURE 18.1 Photo of Mt. Kidd in the Canadian Rocky Mountain Front Ranges, Alberta (Canada). The folds affecting the Paleozoic strata exposed on these cliffs developed in association with transport on the Lewis and Rundle Thrusts. View is to the north.

cleavage. Geologic domains, such as the Canadian Rockies, in which regional horizontal tectonic shortening of the upper-crust yields a distinctive suite of thrust faults, folds, and associated mesoscopic structures, are called **fold-thrust belts** or **fold-and-thrust belts.**

Geologists have been struggling to understand the nature of fold-thrust belts since the 1820s, but it wasn't until the second half of the twentieth century that an integrated image of such belts evolved, because this image could not be developed without data from seismic-reflection profiling (see Chapter 15) and oil-well drilling. Work in the 1960s through 1980s led to a refinement of the geometric and kinematic rules governing the shape and evolution of **thrust-related folds** (folds formed as a result of the development of and slip on thrusts faults), and provided geologists with insight into the fundamental issue of *how* and *why* fold-thrust belts develop. Studies of fold-thrust belts continue in the twenty-first century, as geologists work to characterize fold-thrust belts in three dimensions and try to predict the geometry of their component structures in the subsurface. Current work has practical applications because important oil reserves occur in fold-thrust belts.

In this chapter, for the sake of completeness, we first briefly review the tectonic setting of fold-thrust belts and their regional architecture—some of this material was covered in Chapter 10. We then focus on the geometry of thrust faults and thrust systems, and on the relationships between thrusts and folds. With this background, we can discuss methods for testing the reliability of cross sections depicting fold-thrust belts. We conclude by outlining the key ideas that have been proposed to explain the mechanics by which fold-thrust belts develop. You'll notice that this chapter uses a large number of terms. For better or worse, fold-thrust belt geologists are among the most "jargonistic" folks in geology (see Table 18.1)! Hopefully, you will find that the vocabulary simplifies discussion.

TABLE 18.1 FOLD-THRUST BELT TERMINOLOGY

Allochthon	A mass of rock, comprising a thrust sheet (i.e., a hanging-wall block), that has been displaced by movement on a thrust fault; commonly, use of the term implies that the mass has moved a considerable distance on a detachment from its point of origin.
Allochthonous	Adjective describing "out-of-place" rocks that have moved a large distance from their point of origin.
Autochthonous	Adjective describing rocks that are still at the site where they originally formed and have not been displaced by movement on a thrust fault or detachment.
Backarc	The region that lies behind the volcanic arc along a convergent plate boundary; the backarc and the trench are on opposite sides of the volcanic arc.
Backstop	A representation of the boundary load in the hinterland of a fold-thrust belt. The backstop generates horizontal compressional stress, which contributes to driving fold-thrust belt development. The backstop represents rock of the hinterland that is moving toward the foreland. As such, a backstop is like a snowplow pushing snow toward the foreland.
Backthrust	A thrust on which the transport direction is opposite to the regional transport direction.
Basal detachment	The lowest detachment of a thrust system; the regional basal detachment in a fold-thrust belt separates shortened crust above from unshortened crust below. In the foreland part of a fold-thrust belt, it typically lies at or near the basement-cover contact (also called a basal décollement).
Blind thrust	A thrust that, while it is active, terminates in the subsurface.
Branch line	The line of intersection between two fault surfaces, e.g., where a ramp branches (splays) off of a detachment, or where one ramp splays off another.
Break-forward sequence	A sequence of thrusting during which younger thrusts initiate to the foreland of older thrusts (also called a foreland-breaking sequence).
Break-thrust fold	A fold that initiates prior to thrusting, but later ruptures so that a thrust cuts through its forelimb.
Cutoff (cutoff line)	The line of intersection between a fault and a bedding plane.
Décollement	A subhorizontal fault (also called a detachment)
Detachment	A subhorizontal fault (also called a décollement)
Detachment fold	A fold that forms in response to slip above a subhorizontal fault, much like fold in a rug that wrinkles above a slick floor.
Duplex	A type of thrust system where a series of thrusts branch from a lower detachment to an upper detachment.
Fault-bend fold	A fold that forms in response to movement over bends in a fault surface.
Fault-propagation fold	A fold that forms immediately in advance of a propagating fault tip (also called a **tip fold**).
Floor thrust	The lower detachment of a duplex; it forms the base of the duplex.
Fold nappe	A thrust sheet that contains a regional-scale recumbent fold.
Fold-thrust belt	A geologic terrane in which upper-crustal shortening is accommodated by development of a system of thrust faults and related folds.
Footwall block	The body of rock beneath the fault.
Footwall cutoff	The intersection between bedding planes of footwall strata and a fault surface.
Footwall flat	The portion of the footwall where bedding surfaces parallel the fault.
Footwall ramp	The portion of the footwall where bedding surfaces truncate against the fault (i.e., the portion of the footwall along which there are **footwall cutoffs**).

TABLE 18.1 **FOLD-THRUST BELT TERMINOLOGY**

Forearc	The region to the trench side of the volcanic arc of a convergent plate boundary. The forearc is not the same as the foreland. The forearc lies on the ocean side of a continental volcanic arc.
Foreland	The part of the undeformed craton adjacent to an orogenic belt; some authors have used the term in a more general sense to include the portion of an orogenic belt closer to the undeformed continental interior.
Foreland basin	A sedimentary basin formed on the continent side of a fold-thrust belt that forms because the weight of the stack of thrust sheets in the belt depresses the lithosphere.
Forethrust	A thrust on which the transport direction is the same as the regional transport direction for the whole fold-thrust belt.
Frontal ramp	A ramp that strikes perpendicular to transport direction.
Hanging-wall block	The rock mass that has been transported above a fault surface.
Hanging-wall cutoff	The intersection between bedding planes of hanging-wall strata and the fault surface.
Hanging-wall flat	The portion of the hanging wall where bedding surfaces parallel the fault.
Hanging-wall ramp	The portion of the hanging wall where bedding surfaces truncate against the fault (i.e., the portion of the hanging wall where there are **hanging-wall cutoffs**).
Hinterland	The region closer to the high-grade core of an orogen; as a directional reference, it is the direction opposite to the foreland direction.
Horse	A body of rock in a duplex that is completely enveloped by faults.
Imbricate fan	A type of thrust system where a series of thrusts branch from a lower detachment without merging into an upper detachment horizon.
Inversion tectonics	The process by which a site of extension (e.g., a rift or passive margin basin) transforms into a site of shortening. During inversion, faults that had initiated as normal faults reactivate as thrust faults, and the sedimentary fill of the rift or passive-margin basin is shoved up and over the margins of the basin.
Klippe	An erosional outlier of a thrust sheet that is completely surrounded by footwall rocks; it is an isolated remnant of the hanging-wall block above a thrust.
Lateral ramp	A ramp that strikes parallel to transport direction.
Mechanical stratigraphy	The succession of rock types comprising the stratigraphy of a region, defined in terms of their relative strength.
Oblique ramp	A ramp that strikes oblique to transport direction.
Out-of-sequence thrust	A thrust that initiates to the hinterland of preexisting thrusts.
Out-of-plane strain	The strain due to movement outside the plane of cross section.
Regional transport direction	The dominant direction in which thrust sheets of a thrust belt moved during faulting. Some authors use the term **regional vergence direction** as a synonym.
Roof thrust	The upper detachment of a duplex.
Stair-step geometry	The geometry of a thrust that cuts upsection via a series of flats and ramps. The shape of the fault resembles a staircase in cross section. Typically, the ramps form in stronger units, and the flats in weaker units.
Tear fault	A nearly vertically dipping fault in a thrust sheet that that is parallel or subparallel to the regional transport direction. Motion on a tear fault is dominantly strike-slip and may accommodate differential displacement of one part of a thrust sheet relative to another (i.e., a tear fault is a nearly vertically dipping **oblique ramp** or **lateral ramp**).

TABLE 18.1 FOLD-THRUST BELT TERMINOLOGY

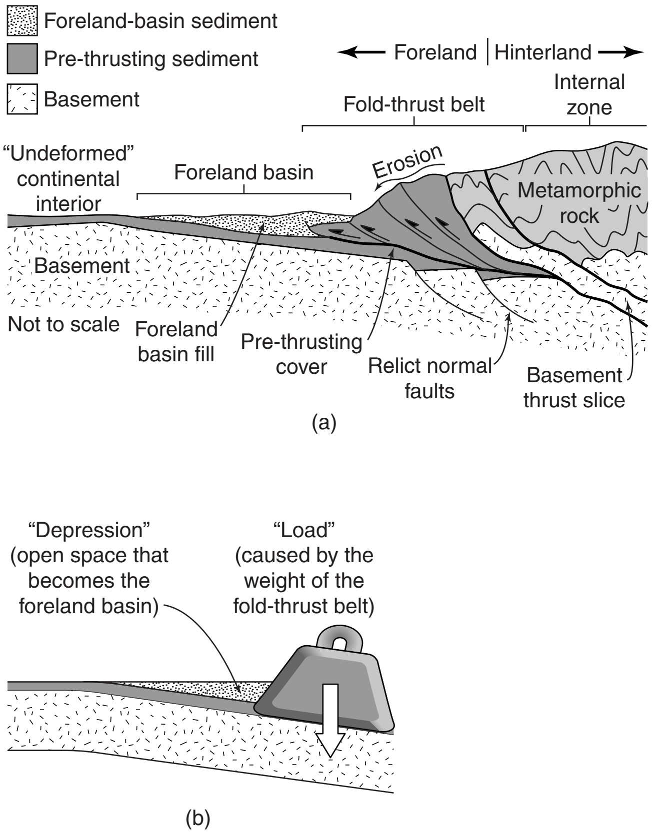
Tectonic inversion	The reactivation of preexisting faults by a reversal of slip direction on the faults.
Thick-skinned tectonics	The process of deformation that involves slip on basement-penetrating reverse faults; this movement uplifts basement and causes monoclinal forced-folds ("drape folds") to develop in the overlying cover.
Thin-skinned tectonics	The process of deformation in which folding and faulting are restricted to rock above a detachment. Some authors restrict the term to situations in which the detachment lies at or above the basement-cover contact. Others use the term even when basement occurs in thrust sheets, to imply that the basement has been transported or detached.
Thrust fault (thrust)	A shallowly to moderately dipping (< 30°) contractional fault with dip-slip reverse movement; in detail, thrusts may include several ramps and flats, and thus on a regional scale, do not necessarily have a uniform dip.
Thrust sheet	The hanging-wall block, above a thrust surface, that has been transported as a consequence of slip on the thrust (also called a **thrust slice**)
Thrust system	An array of related thrusts that connect at depth; a regional-scale thrust system may represent shortening above a specific regional detachment.
Tip line	The line along which displacement on the thrust becomes zero.
Triangle zone	A region in which a wedge of rock is bounded below by a forethrust and is bounded above by a backthrust.
Window (fenster)	An erosional hole through a thrust sheet that exposes the footwall (i.e., an exposure of the footwall completely surrounded by hanging wall rocks).

18.2 FOLD-THRUST BELTS IN A REGIONAL CONTEXT

18.2.1 Tectonic Settings of Fold-Thrust Belts

Fold-thrust belts occur worldwide in a variety of tectonic settings—basically, anywhere that a layer of the upper crust undergoes significant horizontal shortening under low-grade or submetamorphic conditions. To describe these settings, we first need to introduce a few terms. When specifying relative locations in fold-thrust belts that have formed on continental crust, we use the undeformed region of a continent outside of the fold-thrust belt as a point of reference. The **foreland direction** is toward the undeformed continental interior, whereas the **hinterland direction** is toward the orogen's more intensely deformed and metamorphosed internal zone (Figure 18.2a). Similarly, when referring to accretionary prisms formed on ocean lithosphere, geologists sometimes use the term "foreland" to refer to the less deformed side closer to the trench and the term "hinterland" to refer to the more intensely deformed and metamorphosed side of the prism closer to the volcanic arc. Portions of some fold-thrust belts on continents involve strata that had been deposited in a **foreland basin.** In accordance with our definition of "foreland," a foreland basin is a wedge of sediment deposited on the surface of the continent in the foreland of the orogen. Foreland basins form because the stack of thrust sheets in a fold-thrust belt acts as a weight that bends down the surface of the continent and creates a depression that collects sediment eroded from the orogen (Figure 18.2b). Let's now summarize six different settings in which fold-thrust belts develop.

1. *Foreland of an Andean-type convergent margin.* The Andes Orogen is currently developing at a convergent plate boundary, as the floor of the Pacific Ocean slides beneath the continent of South America. Because the South American Plate is moving westward while the Nazca Plate is moving eastward, the two plates squeeze together and crustal shortening takes place across the orogen. This shortening affects the backarc region. (Note that the "backarc region" occurs on the "foreland side" of the orogen—this terminology can be confusing, unless you keep in mind that the term "backarc" specifies



(a)

(b)

FIGURE 18.2 (a) Schematic cross section illustrating the location of a fold-thrust belt in an orogen. Note that the belt occurs between the foreland basin and the internal metamorphic region of the hinterland. Thrusts eventually cut across the strata of the foreland basin and incorporate the basin material into the fold-thrust belt. (b) A stack of thrust slices acts like a heavy load, pushing the surface of the crust down to create a depression that fills with sediment (i.e., to create the foreland basin).

location relative to the volcanic arc, whereas the term "foreland" specifies location relative to the continental interior.) As a consequence, sediment that accumulated along the western margin of South America before convergent tectonism began (Figure 18.3A) as well as strata formed from sediment eroded from the orogen and transported into the foreland basin, undergoes shortening and a fold-thrust belt evolves (Figure 18.3). The Sevier/Laramide fold-thrust belt of the North American Cordillera may be another example of an Andean-type fold-thrust belt.

2. *Accretionary prisms bordering a trench.* We've just focused on what happens in the backarc region of a convergent margin. Don't forget that on the ocean side of the arc, subduction scrapes rock and sediment from the surface of the downgoing plate and incorporates it, along with sediment filling the trench in an accretionary prism (Chapter 17). Accretionary prisms are a type of fold-thrust belt because their development involves shortening and the formation of thrusts and folds (Figure 18.3b). Because the material incorporated in a prism tends to be poorly lithified at the time of incorporation, and because gravity-driven slumping takes place frequently as a prism grows, structures of accretionary prisms tend to be chaotic. Most contemporary examples of accretionary prisms now lie underwater. However, uplifted accretionary prisms do occur in some regions (e.g., Kodiak Island, Alaska, USA).
3. *Foreland sides of a collisional orogenic belt.* Eventually, subduction consumes the oceanic lithosphere between two continents and they collide. When this happens, strata that had originally accumulated in a passive-margin basin[2] on the downgoing plate get caught in a vise between the two continents and undergo tectonic shortening (Figure 18.3c). As a consequence, a fold-thrust belt evolves in which strata of the former passive margin undergo thrusting toward the foreland. During this process, strata of the deeper-water portion of the passive-margin basin may be placed on top of strata of the shallower-water part of the basin. In the hinterland portions of such fold-thrust belts, normal faults formed during the rifting that originally created the passive margin reactivate as thrust faults (i.e., undergo **inversion**), so that basement thrust slices move up and over sedimentary strata. Erosion attacks the rising collisional orogen, providing sediment that collects in a foreland basin. Foreland-basin sediment eventually becomes incorporated in the fold-thrust belt as well. The Valley and Ridge Province of the Appalachians, the Jura Mountains of Switzerland, and the Himalayan Mountains of Asia are examples of fold-thrust belts in collisional orogens.
4. *Inverted rift basins.* We've just described fold-thrust belts formed when an ocean is consumed and continents collide. Similar fold-thrust belts develop when large, sediment-filled rift basins later undergo compression and close. Geologists refer to this process as **rift inversion** to emphasize that a region that was once the site of crustal extension has become the site of crustal shortening.

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[2] Recall that a passive-margin basin forms when rifting succeeds in forming a new ocean basin and the stretched lithosphere along the edges of the continents bordering this ocean slowly cools and subsides. See Chapter 16.

During rift inversion, faults that initiated as normal faults during rifting reactivate as thrust faults (Figure 18.4a and b). Movement on these reverse faults transports the contents of the rift basin up and over the rift's margins. The Pyrenees between Spain and France may represent such a fold-thrust belt.

5. *Seaward edge of passive-margin sedimentary basins.* Not all fold-thrust belts involving strata of passive-margin basins form in regional convergent or collisional settings. While the basin is still growing, gravity may drive the development of thrusts at the seaward edge of the basin, cre-

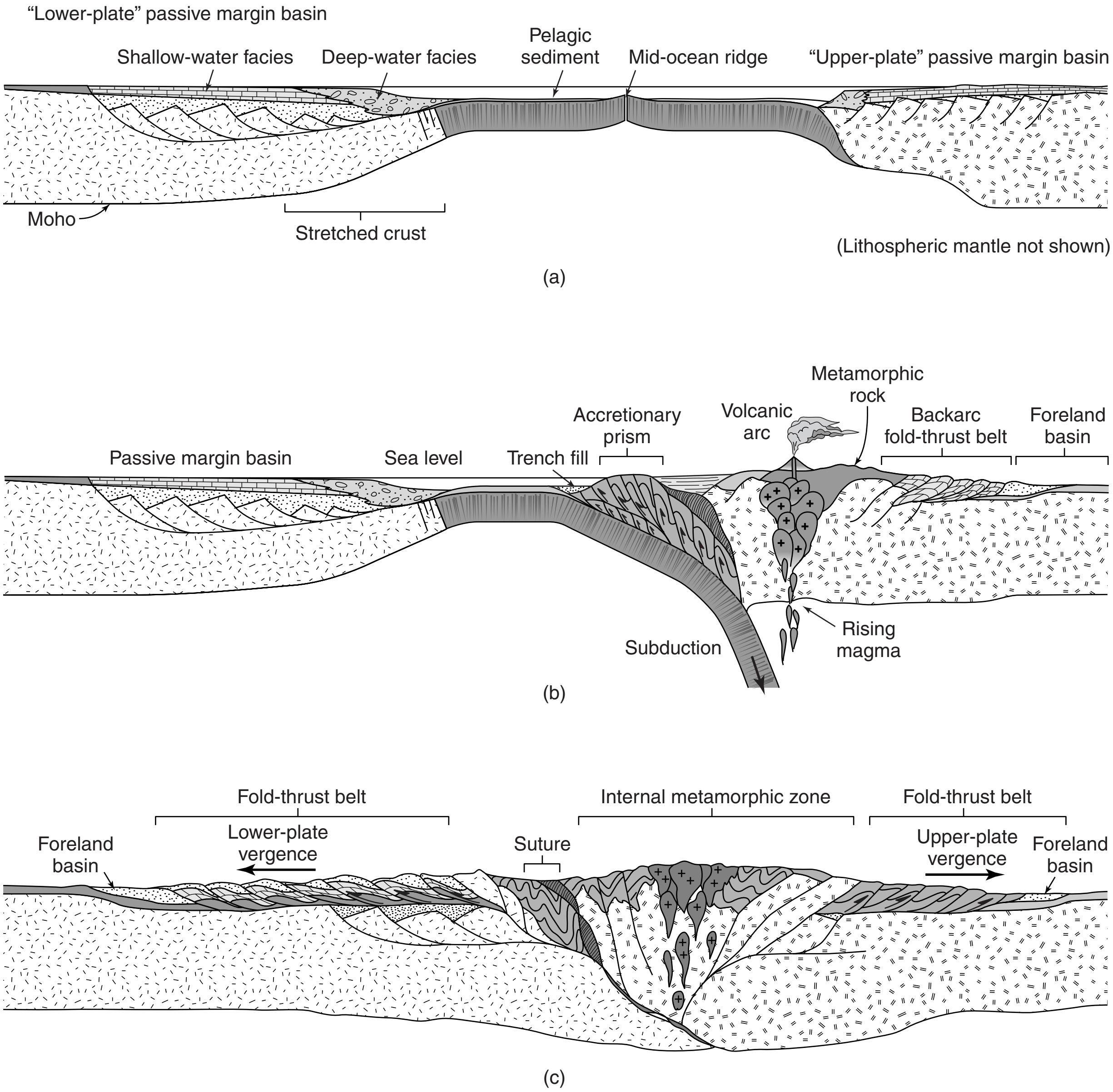


FIGURE 18.3 Regional cross sections depicting stages of fold-thrust belt development first during convergent-margin tectonism and then during continent-continent collision. (a) Passive-margin strata are deposited on thinned continental crust. In this sketch, basins on opposite sides of the margin do not have the same shape, because the basement beneath underwent different amounts of stretching. The so-called lower-plate margin underwent more stretching, whereas the so-called upper-plate margin underwent less stretching. (b) With the onset of convergence, an accretionary prism develops that verges towards the trench, and a backarc fold-thrust belt forms cratonward of the volcanic arc and verges towards the upper-plate craton. (c) Eventually the two continents collide. A fold-thrust belt forms in the foreland of the orogen on both sides of the orogen. Slivers of obducted ocean crust may separate lower-plate rocks from the metamorphic hinterland of the orogen and define the suture between the two plates.

ating an array of structures that resembles a fold-thrust belt. This process happens because gravity causes the strata of passive-margin basins to slump slowly seaward. In the continental shelf region, this movement results in a system of normal faults. But toward the seaward margin of the slumping pile, the mass of sediment slips up and over sediment of the deep ocean floor, just as the downhill edge of a slump on a hillslope rises up and over the ground surface (Figure 18.5a and b). The resulting movement creates a series of thrusts and related folds. Examples of such fold-thrust belts occur along the southern edge of the passive-margin basin of the United States Gulf Coast and along the western coast of Africa.

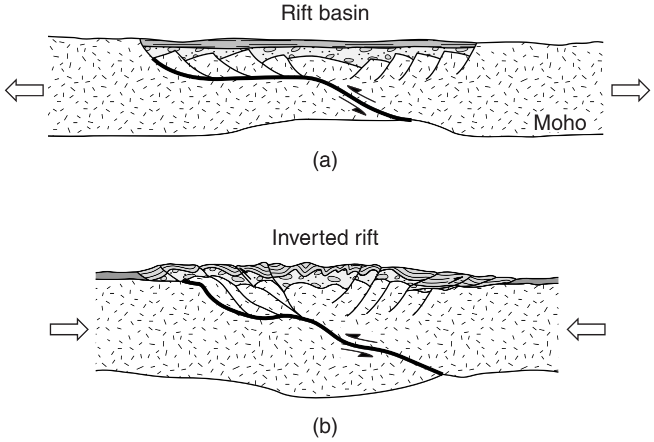


FIGURE 18.4 (a) Cross-sectional sketch of a rift basin just after it has formed. (b) Inversion of the rift occurs when the two margins are pushed towards each other. Note that faults which were originally normal faults turn into thrust faults.

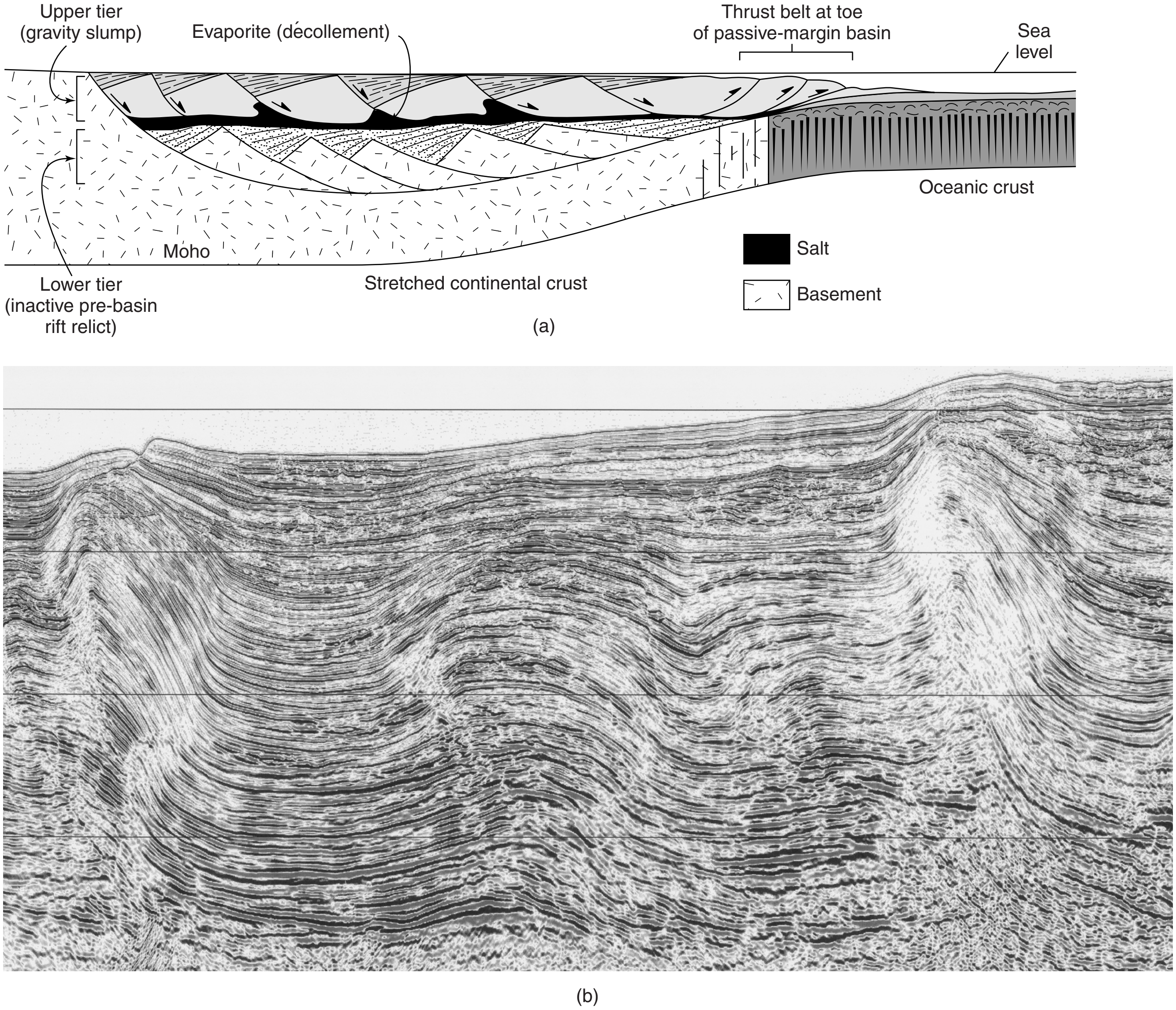


FIGURE 18.5 (a) Cross-section sketch of a fold-thrust belt forming at the seaward toe of a passive-margin basin. (b) Vertically exaggerated two-dimensional seismic-reflection profile illustrating an imbricate fan of thrust faults that has developed offshore of Nigeria.

6. *Restraining bends along large continental strike-slip fault.* Compression develops across a restraining bend along a strike-slip fault. As a result, strata bordering the fault undergo compression and shortening. This deformation may yield a fold-thrust belt bordering the strike-slip fault. Typically, such belts trend oblique to the strike-slip fault. Compression across a restraining bend in southern California has led to the rise of the Transverse Ranges of California adjacent to the San Andreas Fault. In some cases, the fold-thrust belt involves strata that had originally accumulated in a pull-apart basin (i.e., that had undergone transtension) along the fault, so thrusting represents inversion of the pull-apart basin. When this happens, the contents of the basin are thrust out over its former margins (Figure 18.6; see Chapter 19).

18.2.2 Mechanical Stratigraphy

The geometric characteristics of structures in a fold-thrust belt depend on the overall ductility of the rock sequence being deformed and the ductility contrast between layers within the sequence. In other words, fold-thrust belt characteristics depend on the **mechanical stratigraphy**—the succession of strong and weak rock layers—of the sequence being deformed. For example, a sequence consisting of massive layers of limestone behaves differently from one consisting of thin layers of sandstone interbedded with thick layers of shale. The former may break to form several large thrust slices or may flex to form large-amplitude folds, whereas the latter may buckle to form a train of short-wavelength folds. Because different mechanical stratigraphies develop in different tectonic settings, not all fold-thrust belts look the same.

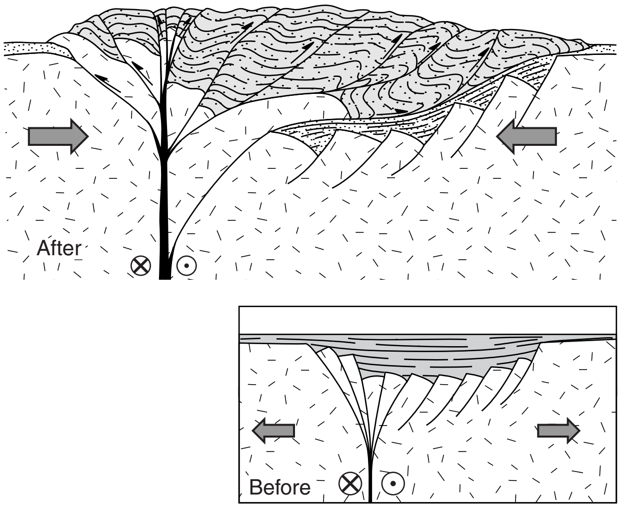


FIGURE 18.6 Cross-sectional sketch illustrating a narrow fold-thrust belt that formed by transpression at a restraining bend along a strike-slip fault. The inset shows the region before transpression began. Here, transtension formed a broad basin prior to inversion.

For much of the remainder of this chapter, we focus discussion on so-called **classic fold-thrust belts.** These belts form where the mechanical stratigraphy subjected to deformation consists of laterally extensive layers that maintain coherence during deformation. In classic fold-thrust belts structures are well organized, in that thrust traces tend to be roughly parallel to one another and folds tend to have roughly the same wavelengths and amplitudes within a given stratigraphic interval. Such mechanical stratigraphy occurs in passive-margin basins and foreland basins. Thus, most of the examples that we use come from collisional and convergent-margin orogens, where deformation involves strata of passive-margin basins and foreland basins.

18.3 GEOMETRY OF THRUSTS AND THRUST SYSTEMS

18.3.1 A Cross-Sectional Image of a Thrust Fault

We've considered the regional setting in which fold-thrust belts occur, now let's focus on the individual structures, and arrays of structures, that occur within fold-thrust belts. To begin our discussion, we examine a cross section of the Pine Mountain Thrust in the southern Appalachians (Figure 18.7a and b). This fault formed in a sequence of Paleozoic strata during the Alleghanian Orogeny, when Africa collided with North America.

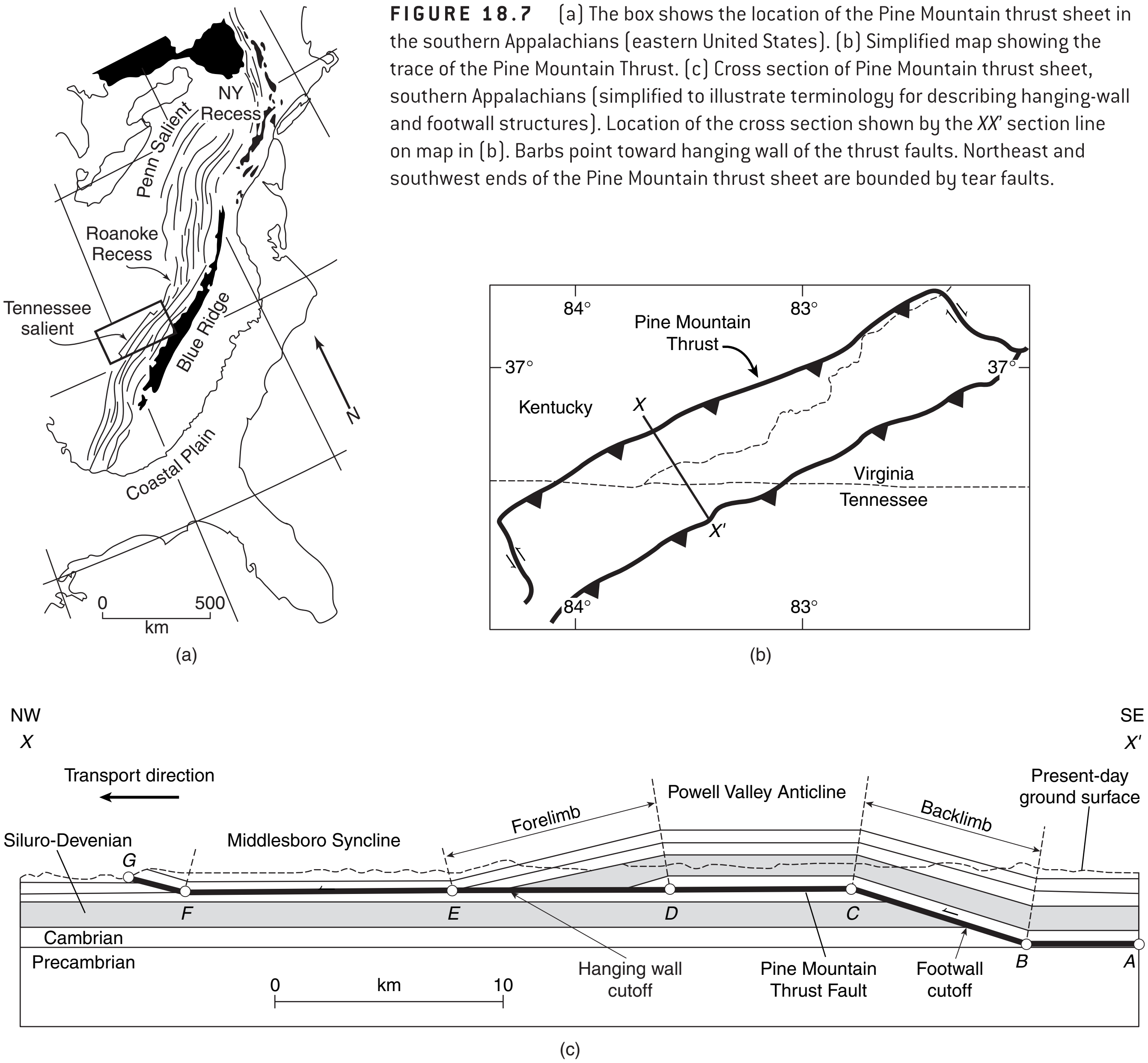
As shown in Figure 18.7c, thrust faults like the Pine Mountain Thrust cut *upsection* in the direction that the hanging-wall moves. Displacement on the fault puts older strata on top of younger strata. In the case of the Pine Mountain Thrust, we see that the fault lies at the base of the Cambrian strata in the southeast, cuts upsection to the northwest, and eventually flattens out in Siluro-Devonian strata. Movement on the Pine Mountain Fault placed Cambrian strata over Siluro-Devonian strata. Note that thrusting can duplicate a stratigraphic succession, so that a vertical hole drilled through the hanging wall, across the fault, and into the footwall could encounter the same stratigraphic units twice. Further, thrusting raises strata above its pre-faulting elevation. Strata in the hanging wall of the

Pine Mountain Thrust lie approximately 2.5 km above their original pre-faulting elevation!

Our example of the Pine Mountain Thrust also illustrates that some thrust faults resemble a flight of stairs, in that they consist of **flats** that lie approximately in the plane of bedding and **ramps** that cut across bedding (Figure 18.7c). The key to determining whether a fault segment is a ramp or a flat is to look for cutoffs. A **cutoff** is the intersection between a bedding plane and a fault surface along a ramp. Flats commonly exceed ramps in cross-sectional length, and typically lie within incompetent (weak) strata like shale and evaporite. Ramps tend to develop in competent (strong) rocks like sandstone, dolomite, and limestone. Note that their placement depends on mechanical stratigraphy.

If you examine Figure 18.7c in detail, you will see that a ramp with respect to the footwall can be a flat with respect to the hanging wall. Thus, to be complete, a description of a given ramp should indicate whether it is a **hanging-wall ramp** that cuts across beds of the hanging wall, or a **footwall ramp** that cuts across beds of the footwall. Similarly, a description of a flat should indicate whether the flat is a **hanging-wall flat** that lies parallel to bedding in the hanging wall, or a **footwall flat** that lies parallel to bedding in the footwall. Note that some segments of a fault may display a "flat-on-flat relationship," meaning that the segment is a flat

FIGURE 18.7 (a) The box shows the location of the Pine Mountain thrust sheet in the southern Appalachians (eastern United States). (b) Simplified map showing the trace of the Pine Mountain Thrust. (c) Cross section of Pine Mountain thrust sheet, southern Appalachians (simplified to illustrate terminology for describing hanging-wall and footwall structures). Location of the cross section shown by the *XX'* section line on map in (b). Barbs point toward hanging wall of the thrust faults. Northeast and southwest ends of the Pine Mountain thrust sheet are bounded by tear faults.



with respect to both the hanging wall and footwall. In Figure 18.7c, segment *FE* is a hanging-wall flat on a footwall flat, *ED* is a hanging-wall ramp on a footwall flat, and *CB* is a hanging-wall flat on a footwall ramp.

To better understand the cross-sectional geometry of a thrust, let's examine the relationships among ramps and flats on a fault both before and after slip. Figure 18.8a shows the trace of a stair-step thrust fault before movement has taken place. Before movement, hanging-wall ramps must lie adjacent to footwall ramps, and hanging-wall flats must lie adjacent to footwall flats. Thus, the number of hanging-wall flats and ramps *exactly* matches the number of footwall flats and ramps. After slip (Figure 18.8b), the hanging-wall block moves, so a hanging-wall ramp may end up on a footwall flat, and a hanging-wall flat may end up on a footwall ramp. Before slip on the fault, the hanging-wall ramp between *T* and *U* was adjacent to the footwall ramp between *V* and *W*, and the hanging-wall ramp between *P* and *Q* was adjacent to the footwall ramp between *R* and *S*. Note that after slip, the number of hanging-wall flats and ramps still matches the number of footwall flats and ramps.

Typically, ramps curve and change strike along their length. A ramp segment that strikes approximately perpendicular to the direction in which the thrust sheet moves is a **frontal ramp** (Figure 18.9), a ramp segment that cuts upsection laterally and strikes approximately parallel to the direction in which the thrust sheet moves is a **lateral ramp,** and a ramp segment that strikes at an acute angle to the transport direction is an **oblique ramp.** Note that dip-slip movement dominates on frontal ramps, oblique-slip movement dominates on oblique ramps, and strike-slip movement dominates on lateral ramps. **Tear faults** are lateral or oblique ramps that break a thrust sheet into segments that move by different amounts into the foreland.

While active, some thrust faults lie entirely in the subsurface, whereas others cut and displace the ground surface during movement. Consequently, geologists distinguish between **blind thrusts,**[3] which are faults that moved in the subsurface and do not intersect the "syntectonic ground surface" (the ground surface at the time of active faulting), and **emergent thrusts,** which are faults that intersected and displaced the syntectonic ground surface. After exhumation of a thrusted region by uplift and erosion, it is not always possible to determine if a given fault was emergent or blind while it was active.

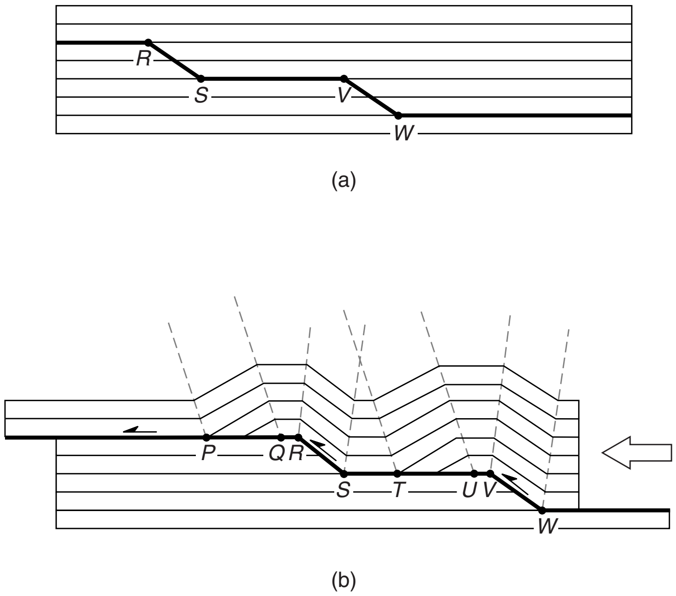


FIGURE 18.8 When a thrust sheet moves up a stair-step fault, ramp anticlines develop. (a) The cross-sectional trace of the fault before slip. Points at the tops and bottoms of ramps are labeled. (b) The cross-sectional trace of the fault after slip. Note that the number of hanging-wall ramps exactly matches the number of footwall ramps.

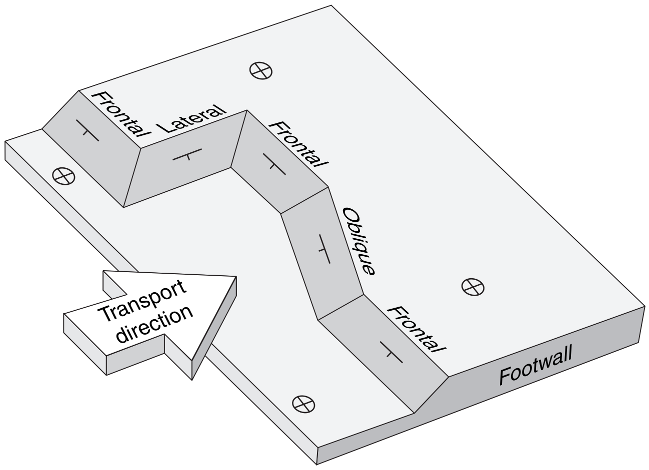


FIGURE 18.9 Three-dimensional block diagram illustrating different types of fault ramps (hanging wall removed). Tear faults are vertically dipping oblique ramps or lateral ramps.

If erosion produces a hole in a thrust sheet so that at the ground surface we see an exposure of the footwall completely surrounded by hanging-wall rocks, then the exposure is called a **window** or **fenster** (Figure 18.10a). Alternatively, if erosion removes most of a thrust sheet so that you can map a remnant of the thrust sheet that is completely surrounded by footwall strata, then the "island" of hanging-wall rock is called a **klippe** (Figure 18.10b).

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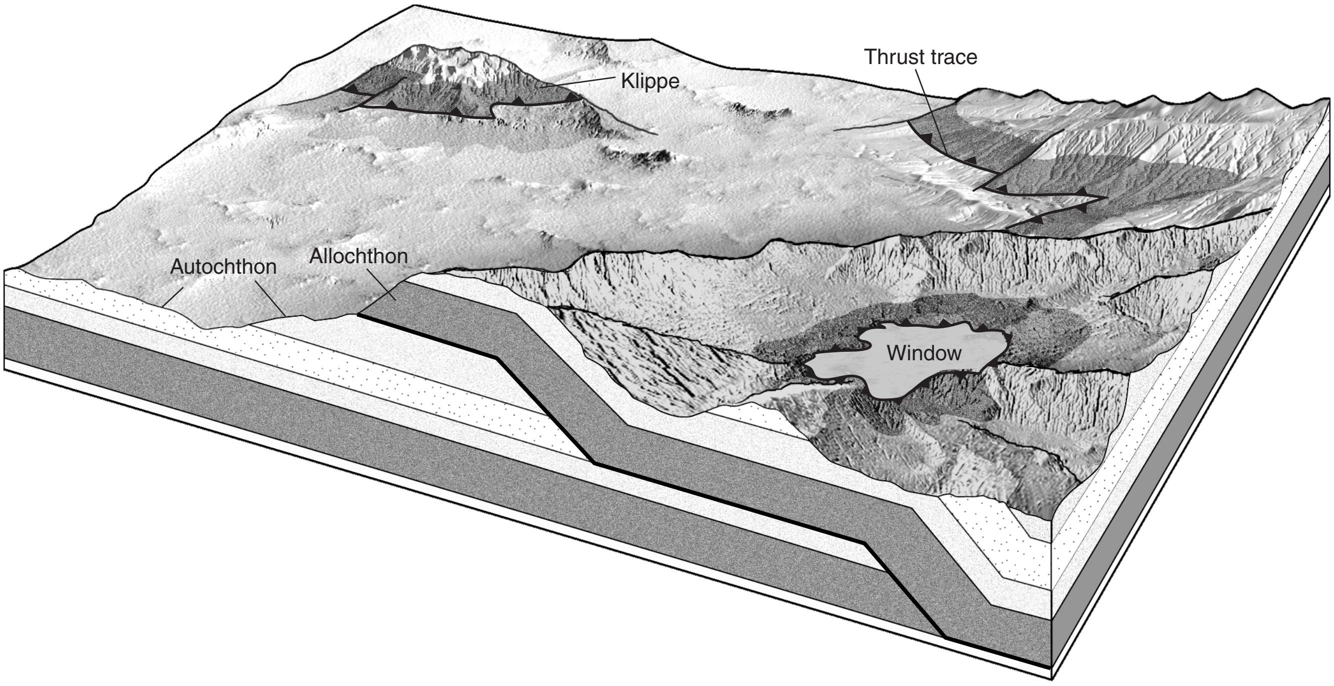
[3] Notably, the disastrous 1994 Northridge earthquake in California occurred on a blind thrust fault; this fault had not been previously identified by any surface expression, and thus movement on it was a surprise.



(a)

(b)

FIGURE 18.10 (a) Block diagram illustrating klippe, window, allochthon, and autochthon. (b) Photo of Crow's Nest Mountain Klippe, Alberta (Canada).

18.3.2 Thrust Systems

A **thrust system** refers to the family of related thrust faults that ramp up from a single **detachment fault** or **décollement** (e.g., Figure 18.11). The name "detachment" emphasizes that rock above the fault has detached or separated from rock below during movement. Detachments tend to develop in weak rock types, such as shale or evaporite.

There are two end-member types of thrust systems—imbricate fans and duplexes. Individual thrusts that make up an **imbricate fan** branch upsection from a common detachment and terminate updip without merging into an upper detachment (Figure 18.12a). The line (in three dimensions) along which the ramp connects to an underlying detachment is called a **branch line,**[4] and the line at which the fault terminates and displacement decreases to zero is called the **tip line.** In a **duplex,** a series of thrusts branches upwards from a lower detachment and merges with a higher detachment (Figure 18.13). Sometimes geologists refer to the lower detachment of a duplex as the **floor thrust,** and the upper one as the **roof thrust** (Figure 18.13a and b). Note that adjacent thrust surfaces in a duplex completely surround bodies of rock; these fault-bounded bodies are called **horses.**[5] Depending on the spacing and relative displacement on thrusts in a duplex, the roof thrust of a duplex may be planar (Figure 18.13a) or corrugated (Figure 18.13c). The roof thrust and overlying strata of a corrugated-roof duplex is folded into a train of anticlines and synclines.

In both imbricate fans and duplexes, the faults comprising a thrust system do not all initiate at the same time. Generally they initiate in a **break-forward**

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[4] The term "branch line" can also be used to define the line at which a ramp bifurcates to form two separate ramps, either updip or along strike.

[5] Note that a "horse" is not the same structure as a "horst."

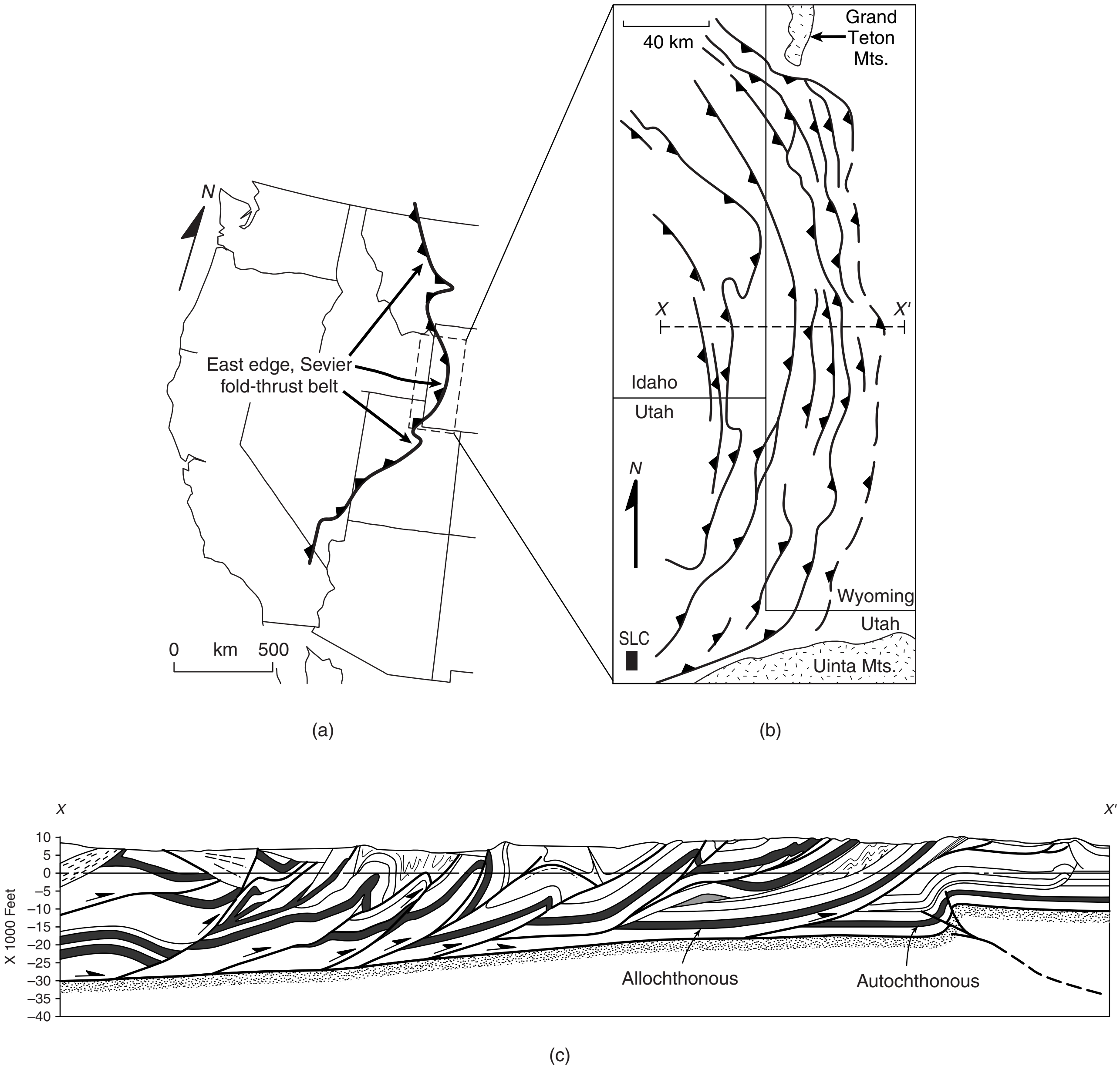

FIGURE 18.11 (a) Trace of the eastern edge of the Sevier fold-thrust belt in the western United States. (b) Map showing the traces of principal thrusts in the western Wyoming fold-thrust belt (a part of the Sevier fold-thrust belt). Note the thrusts curve so they are convex toward the foreland. (c) Cross section of the western Wyoming fold thrust belt. Approximate location of the cross section is indicated by the *XX*' section line in (b). (c) Cross section of the western Wyoming fold thrust belt.

sequence (Figure 18.12). This means that the faults of the system form one after the other, with each new fault forming to the foreland side of the previous one. Thus, the youngest fault in the system occurs on the foreland end of the system, whereas the oldest fault occurs at the hinterland end. There are exceptions to the break-forward sequence model. Slip at any given time may be partitioned among the youngest thrust and thrusts immediately behind it. Also, in some cases, existing thrusts of the system reactivate and/or new faults initiate to the hinterland of the preexisting faults. These **out-of-sequence faults** can be recognized where they cross cut structures in the foreland.

Within a given fold-thrust belt, most thrust sheets move in the same overall direction, called the **regional transport direction** (Figures 18.11, 18.12, and 18.13).[6] In the case of classic collision-related or

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[6] Some geologists also use the term "vergence" to describe transport of thrust sheets as in, "Thrust sheets of the Ouachita Mountains verge to the north."

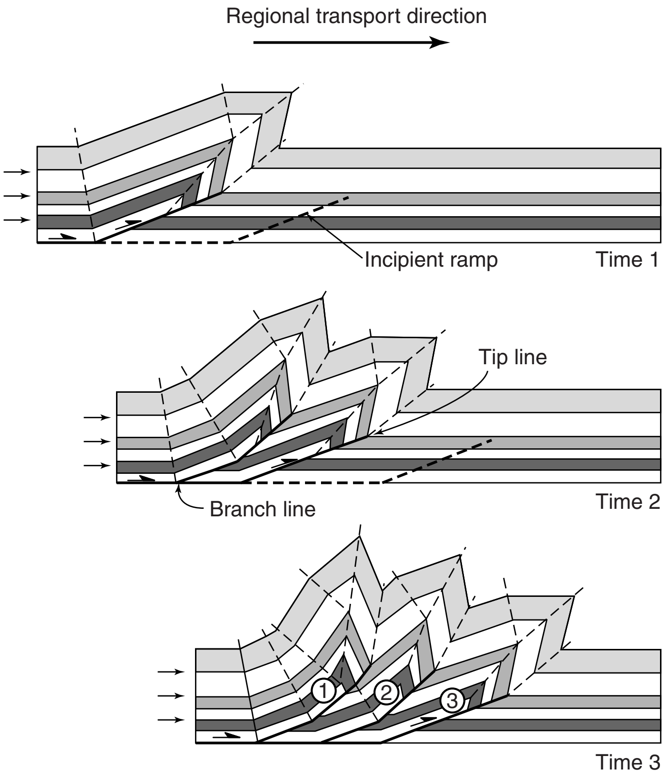

FIGURE 18.12 An idealized imbricate fan that develops by progressive break-forward thrusting. Note that successively younger thrusts cut into the footwall, and older faults and folds become deformed by younger structures. The dashed lines are the traces of fold axial surfaces. In the cross section showing "Time 3," the sequence of thrusts is labeled. Fault 1 is the oldest and Fault 3 is the youngest. On this cross section, tip lines and branch lines are points; in three dimensions, they go into and out of the page.

convergent-margin fold-thrust belts, the regional transport direction carries rocks from the hinterland of an orogen toward the foreland. For example, the regional transport direction in the Appalachian fold-thrust belt is to the west, whereas the regional transport direction for the Canadian Rockies is to the east. Regional transport direction in accretionary prisms is generally toward the trench.[7] In discussion, it is convenient to use the term "vergence" to describe the movement of thrust sheets. For example, if the regional transport direction in a thrust system is to the east, we could also say that the thrusts "verge to the east."

In detail, however, not all thrust sheets of a fold-thrust belt move in the same direction, so we distinguish between **forethrusts** that verge in the regional transport direction and **backthrusts** that verge opposite to the regional transport direction. Backthrusts may form where the front of a thrust sheet wedges between layers of strata in the foreland as it moves up and over a footwall ramp and onto a footwall flat, or in the hanging wall as a thrust sheet rides up over a ramp (Figure 18.14a). A place where the thrust sheet wedges between layers is called a **triangle zone** (Figure 18.14b).

So far, we've focused on faults that branch from a detachment. It's important to point out that not all thrusts do so. Local thrusts may form during evolution of a fold-thrust belt, when synclines tighten so that their limbs move toward each other. Because of this tightening, there is no longer enough room in the hinge zone for the volume of rock within the fold. When such "room problems" develop, **out-of-the syncline faults** form to transport rock up and out of the hinge zone (Figure 18.14c). Note that out-of-the-syncline faults can be either forethrusts or backthrusts and that displacement decreases downdip, so that the faults die out within the fold.

18.3.3 Overall Fold-Thrust Belt Architecture

Let's now broaden our view to incorporate an entire classic fold-thrust belt. For the sake of discussion, we divide the belt into two parts, a foreland part and a hinterland part.

In the foreland portion of a classic fold-thrust belt, deformation is restricted to the rock above a regional fault called the **basal detachment** or **basal décollement** (Figure 18.11c). A basal detachment is the lowest detachment of a fold-thrust belt. The belt may contain higher level detachments and many imbricate fans and duplexes, as well as several detachment horizons, but they all lie above the basal detachment. Typically, the basal detachment of the foreland part of a fold-thrust belt lies in a weak shale or evaporite at or near the basement-cover contact.[8] Toward the foreland, the basal detachment may ramp upsection to higher stratigraphic levels. Rocks above the detachment are **allochthonous** in that they have been transported relative to their original location. In contrast, rocks that lie

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[7] Some accretionary prisms, however, are bivergent, meaning that thrust sheets on the trench side of the prism move toward the trench, while those on the arc side move toward the arc.

[8] Commonly the "basement" consists of older crystalline rocks (e.g., Precambrian gneiss and granite) that form the substrate on which sedimentary beds were unconformably deposited. The term "cover" refers to the succession of overlying sedimentary beds.

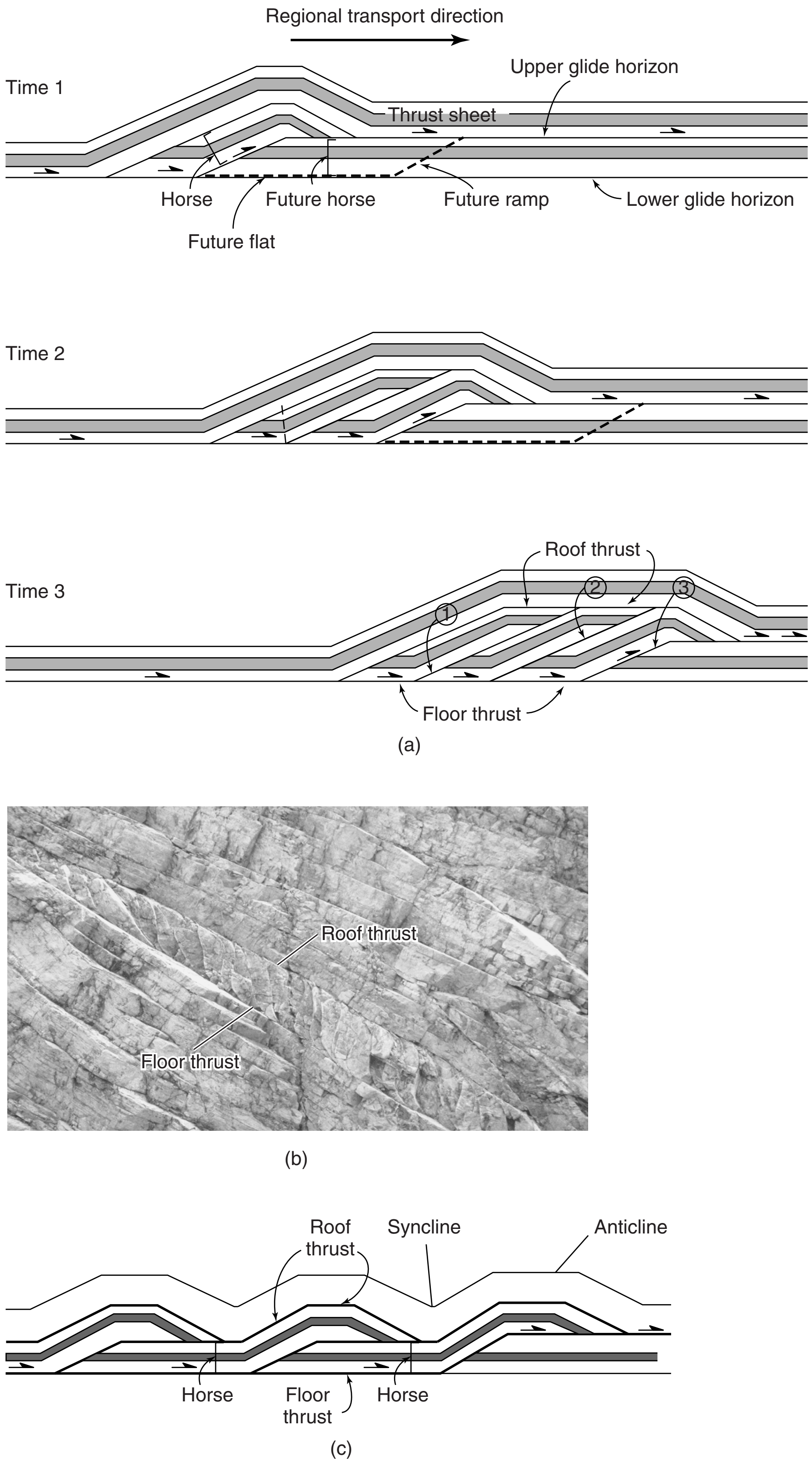

FIGURE 18.13 (a) Idealized flat-roofed duplex that develops by progressive break-forward faulting. Note that the roof thrust undergoes a sequence of folding and unfolding, and that formation of the duplex results in significant shortening. (b) Photo of a duplex involving a single bed at Crow's Nest Pass, Alberta (Canada). Duplex height is about 0.3 m. (c) Schematic sketch, using kink-style fold construction, of a corrugated roof duplex.

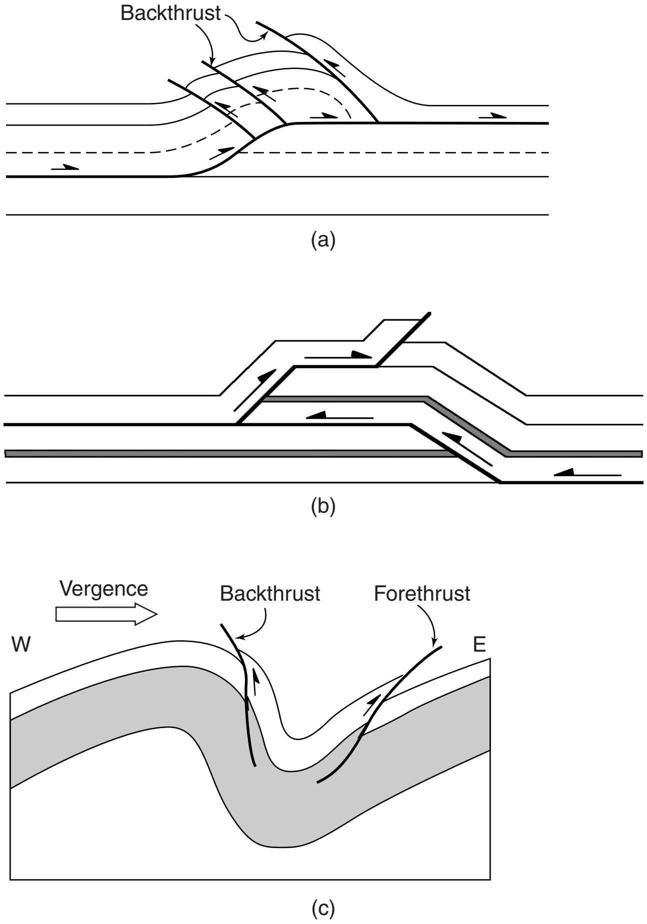

FIGURE 18.14 (a) Formation of backthrusts across the backlimb of a ramp anticline, and along bedding of the forelimb. (b) Cross section of a triangle zone. (c) Out-of-the-syncline forethrusts and backthrusts. These faults formed to accommodate a room problem that develops in the hinge zone of the syncline.

below the detachment are **autochthonous,** in that they have not been transported by fault slip and thus lie in their original position.

In the hinterland portion of a fold-thrust belt, the basal detachment "roots" into basement (Figure 18.3b and c). By this we mean that basement rocks are incorporated in thrust sheets. In some cases, the basement slices have been transported long distances and end up on top of sedimentary rock. Basement slices in the hinterland of fold-thrust belts probably originated as the hanging-wall blocks above normal faults that formed during the rifting stage, prior to passive-margin basin formation. Thus, thrusting of these slices represents inversion of the rift that underlies the passive-margin basin. Further into the hinterland, deformation occurs under metamorphic conditions, so sedimentary rocks are able to deform plastically. Sedimentary layers incorporated in hinterland thrust slices may be folded into huge recumbent folds, underlain by a detachment. Such folds are called **fold nappes.** The Swiss Alps contain excellent examples of fold nappes (Figure 18.15; see Section 21.1).

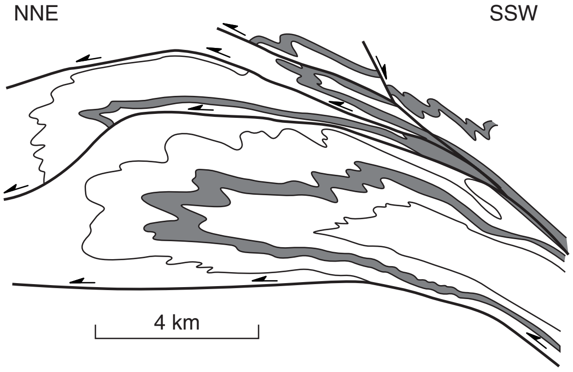

FIGURE 18.15 Profile sketch (based on down-plunge projection) of the Helvetic nappes in the Alps. Note that the limbs of these recumbent folds have been thinned during shearing.

Geologists sometimes refer to the style of deformation in which faulting and folding occur only above a regional basal detachment as **thin-skinned tectonics.** Unfortunately, this term has not been used consistently in the literature. Some researchers use it to distinguish faulting that does not involve basement from faulting that does, and use the term "thick-skinned deformation" for the latter.[9] Others use the adjective "thin-skinned" in all cases where the deformed interval is underlain by a subhorizontal detachment, regardless of whether or not basement is involved. In general, the adjective "thick-skinned" is preferred for the style of deformation in which movement occurs on basement-penetrating reverse faults, which have fairly steep dips near the ground surface. These faults may die out up-dip in a monocline.

18.4 THRUST-RELATED FOLDING

Geologists refer to the Canadian Rockies and the Appalachian Valley and Ridge Province as "*fold*-thrust belts" because, in addition to the thrust systems that we have just described, these regions contain spectacular folds, with amplitudes ranging from millimeters up to a few kilometers (Figure 18.16a and b). We call these folds "thrust-related folds" because they form in

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[9] As alternative terminology, one could distinguish between "basement-detached" deformation and "basement-involved" deformation. In the former, the detachment lies at or above the basement-cover contact, while in the later, basement rocks have been incorporated in thrust slices.

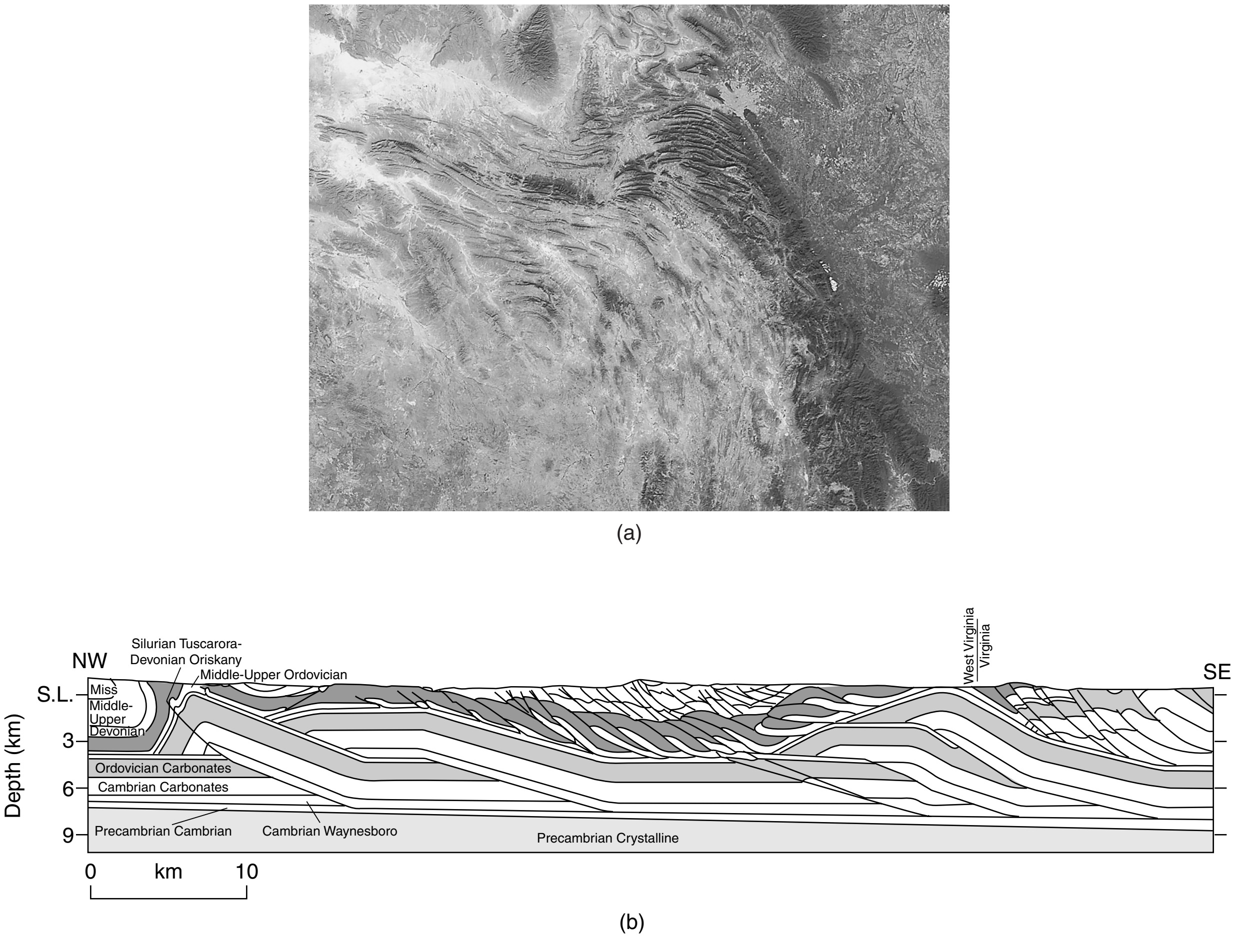

FIGURE 18.16 (a) Space photo of the Monterray Salient, Mexico, a prominent fold belt comprised predominantly of detached folds in Cretaceous carbonates. Note that the fold belt transitions to a fold-thrust. The features to the foreland (top of the photo) are salt diapirs. (b) Cross section of the Appalachian Valley and Ridge Province in Virginia and West Virginia. Observe that there are duplexes at two levels; the roof thrust of the Cambrian-Ordovician duplex is corrugated and is the floor thrust for a higher-level duplex.

association with displacement on thrust faults. We recognize four broad categories, based on the specific relationship between a fold and the underlying fault:

1. *Folding associated with break thrusts.* Picture the development of a fold resulting from the buckling of a stratified sequence. Initially, the layers bend, without rupturing, to form an open anticline-syncline pair (Figure 18.17a). The folds tighten and become more asymmetric as folding progresses. Eventually, strain can no longer be accommodated by folding alone, and rupturing produces a thrust fault that breaks through the overturning limb (the "forelimb") of the anticline. Because the thrust cuts through the limb of an already formed fold, it is called **break thrust**[10] (Figure 18.17b). Note that a break thrust develops *after* folding. After a break thrust has ruptured the forelimb of an asymmetric anticline, it displaces the anticline of the hanging wall to the foreland, relative to the syncline of the footwall.
2. *Fault-bend folding:* Picture a simple thrust geometry in which a thrust that occurs as a flat in a weak layer bends, cuts upsection across a strong layer as a ramp, and then bends again and becomes a flat at the top of the strong layer (Figure 18.18a-b). If you push on the thrust sheet,

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[10] Admittedly, this isn't an ideal name, as the term itself does not immediately bring to mind the process we've described, but researchers have ensconced the term in fold-thrust belt jargon.

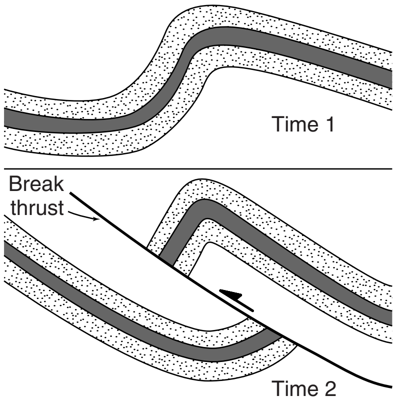

FIGURE 18.17 Formation of a "break-thrust fold." (a) An asymmetric fold begins to form and tighten. (b) Eventually, a fault breaks through the fold's forelimb.

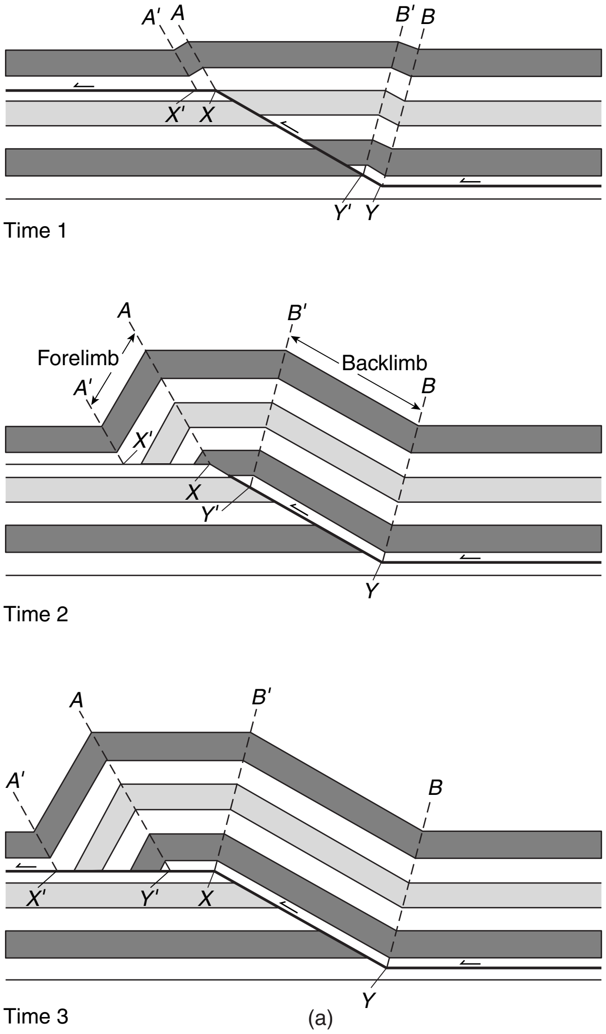

FIGURE 18.18 (a) Cross-sectional model showing the progressive stages during the development of a fault-bend fold. The dashed lines are the traces of axial surfaces. (b) Photo of a fault-bend fold above the McConnell Thrust, near Seebe, Alberta (Canada). These Paleozoic strata have been displaced over 5 km vertically and 40 km horizontally, and now lie above Cretaceous foreland basin deposits. The foreland basin deposits have preferentially eroded and are now forested.


(b)

the sheet must itself bend to climb the ramp. Because of Earth's gravity, the sheet cannot rise into the air at the top of the ramp. Rather, the layer bends again to form an anticline on the upper flat. Thus, by pushing strata up and over a preexisting stair-step thrust, strata of the thrust sheet must undergo folding. Note that this folding occurs *after* the formation of the thrust. The resulting fold is called a **fault-bend fold.**[11] In sum, a fault-bend fold forms where hanging-wall strata move up and over a stair-step in a fault; the strata deform in order to conform to changes in dip (bends) of the fault surface.

The geometry of a specific fault-bend fold depends on the geometry of the stair-step thrust beneath it. Note that if the bends in the fault surface are abrupt, the fold will have kink-style (i.e., not rounded) hinges. Also, note that once the fold has climbed over the ramp, the width of the anticline depends on the magnitude of displacement on the fault; as displacement increases, interlimb distance increases because the back limb (the limb closer to the hinterland) remains over the ramp (Figure 18.19). In an ideal fault-bend anticline, the fold's backlimb parallels the footwall ramp, the fold's forelimb is shorter and steeper than the backlimb, strata of the footwall remain flat-lying, and the kink-style hinges of the hanging-wall anticline directly reflect the shape of bends in the fault surface. In some cases, strata move through the kink hinges as the fold evolves.

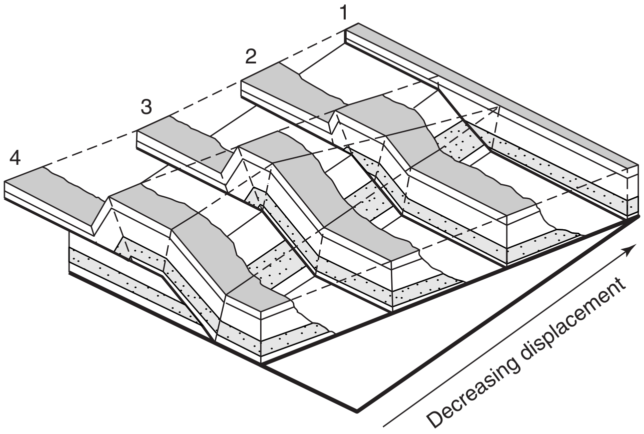

FIGURE 18.19 Cutaway block diagram depicting a possible along-strike termination geometry of a simple fault-bend fold that experiences an along-strike decrease in displacement, from a maximum on section 4, to zero on section 1. The dashed lines are the traces of axial surfaces, and the dotted lines are position reference lines.

3. *Fault-propagation folding.* In some cases, folding develops just in advance of the tip of a ramp as the ramp propagates updip. The resulting fold is called a **fault-propagation fold.** Such folds develop *concurrently* with thrust development. Typically, a fault-propagation fold develops only in front of the upper tip line as the ramp propagates upsection toward the ground surface. Fault-propagation folds are asymmetric and verge in the direction of thrusting (Figure 18.20a and b). Displacement along the fault dies out in the hinge zone of the fold.

   In a geometric model of fault-propagation folds (Figure 18.20), the fault tip propagates upsection as the fold's backlimb and forelimb lengthen. The fault tip and the merge point of the fold's kink axes lie at the same stratigraphic horizon. As in the fault-bend fold model, the backlimb of a fault-propagation fold parallels the dip of the ramp, and strata in the footwall remain flat-lying. In some cases, strata move through the kink axes as the fold evolves.

   In recent years, geologists have examined deformation in the region above the tip line of the fault, and have found that not all regions obey the classical geometric image of a fault-propagation fold with kink-style hinges as described above. Rather, a triangular (in profile) region of deformation develops beyond the fault tip. This region has come to be known as a **trishear** zone.[12] Figure 18.21 shows the geometric characteristics of a trishear deformation zone. You will note that because strain is distributed through the triangular region, folds have curved hinges, and beds undergo stretching and thinning.

4. *Detachment folding.* Folds may develop in fold-thrust belts above a detachment fault, even if no ramps develop. This happens where the strata above the detachment buckle or wrinkle up like a rug that has been shoved across a slick wooden

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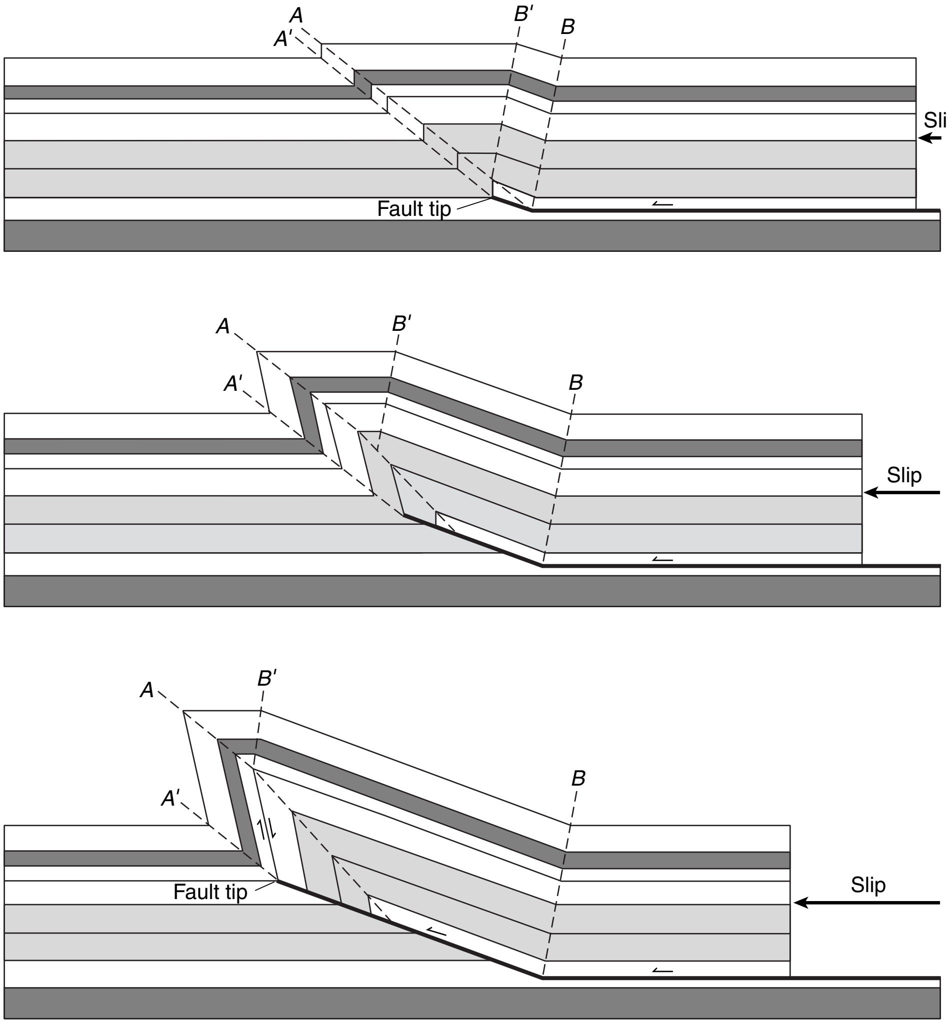
[11] The geometry of fault bend folds was first worked out rigorously by John Suppe of Princeton University, in classic papers published in the early 1980s.

[12] The concept of trishear deformation was proposed by Eric Erslev of Colorado State University. Richard Allmendinger of Cornell University has developed insightful computer models of trishear deformation.



(a)

FIGURE 18.20 (a) Model for progressive development of a simple fault-propagation fold. (b) Exposure of a fold in the Lost River Range, Idaho, showing an asymmetric fold dying out updip in the core of a fold.

(b)

floor (Figure 18.22). Such **detachment folds** are particularly common in regions where detachments lie within thick shale or salt layers, as occurs in the Jura Mountains of Switzerland, for the weak rock can flow into the core of the fold as the structure develops. In some cases, a break thrust may develop at a late stage in the evolution of the fold; the fault cuts across the forelimb of the already formed detachment fold.

Division of folds in fold-thrust belts into the classes listed above is clearly an oversimplification. Today, geologists realize that an evolutionary continuum exists between these classes. For example, initial detachment folding may establish the spacing for folds. Further shortening causes fold amplification and tightening, with the initiation and propagation of a break thrust further modifying the overall fold geometry. If displacement along the fault is sufficiently high, the fault may break through the fold forelimb, transporting the fold along the fault. If the thrust merges updip with a flat, then continued displacement of the thrust sheet will cause the fold to evolve into a fault-bend fold. Alternatively, a thrust may "lock up" after some displacement has occurred. When this happens, the fault plane ceases to grow, and shortening in the area is accommodated by folding rather than by frictional sliding.

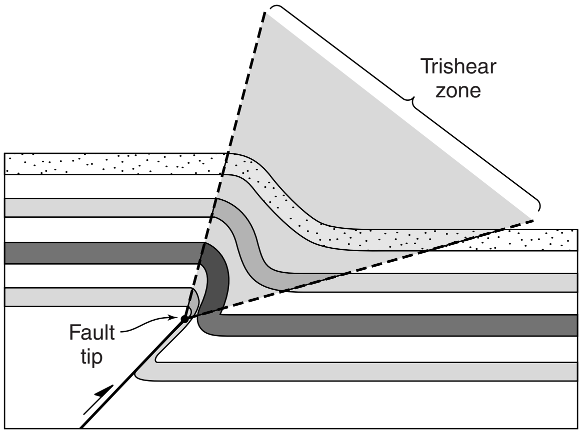


FIGURE 18.21 Cross section illustrating the concept of trishear deformation. Note that strain is distributed throughout a triangular zone in the region beyond the fault tip. The solid line is the fault trace and the dashed lines outline the region of trishear.

Before leaving our discussion of folds, we point out that the folds we've described above can be considered to be the "first-order folds" of fold-thrust belts (see Chapter 10). Typically, thrusting and folding also produce "second-order folds" as well, meaning smaller folds that form within larger folds. Folds in this category include parasitic intraformational folds formed due to flexural slip between beds on the limbs of larger folds, folds that develop within shear zones, and folds formed by buckling of thinner-bedded intervals in the hinge zones of larger folds.

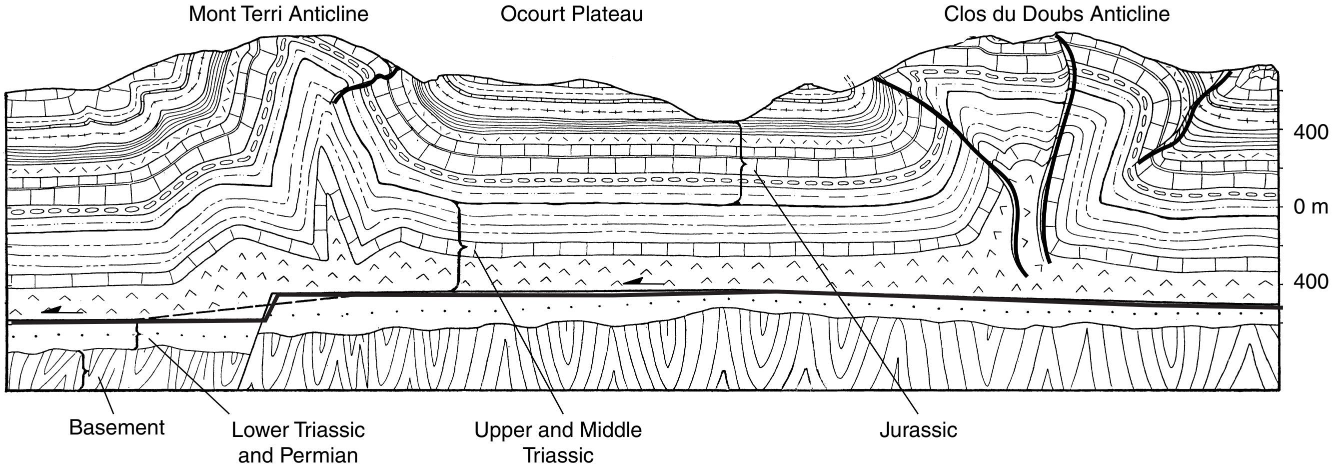


FIGURE 18.22 Cross section of detachment folds in the foreland of the Jura Mountains, Switzerland. Note that the folds do not involve the basement. The fold cores filled with ductile rock as the folds formed.

18.5 MESOSCOPIC- AND MICROSCOPIC-SCALE STRAIN IN THRUST SHEETS

Fold-thrust belts form in response to layer-parallel compression of the upper crust. This means that during the development of a fold-thrust belt, the maximum principal compressive stress (σ_I) is horizontal and has a bearing roughly perpendicular to map trace of folds and faults within the fold-thrust belt.[13] If you have a chance to study a fold-thrust belt up close, you'll discover that folds and thrusts are not the only structures that they contain. Compression also may cause a suite of mesoscopic- and microscopic-scale structures to form in fold thrust belts. Specific examples include the following:

1. *Tectonic cleavage.* Appropriate rock types (e.g., shale, argillaceous limestone, and argillaceous sandstone) tend to be susceptible to pressure-solution deformation and thus develop tectonic cleavage (Figure 18.23a-b). This cleavage varies from widely spaced cleavage to slaty cleavage. Generally, the strike of cleavage is approximately parallel to the trends of folds. Dip of the cleavage varies; slaty cleavage tends to be parallel to the axial planes of folds, whereas spaced cleavage tends to fan around folds. Adjacent to faults, cleavage becomes inclined at a low angle to the fault surface, for the cleavage domains rotate toward the direction of transport.
2. *Mesoscopic folds.* Stratigraphic units that consist of thin beds of relatively strong rock types interbedded with layers of relatively weak rock types may undergo buckling during the shortening that produces the fold-thrust belt. As a result, mesoscopic folds may develop within some thrust sheets (Figure 18.23a).
3. *Wedge thrusts.* If a thrust sheet contains a succession of strata in which strong beds are interlayered with weak beds, on the scale of centimeters to meters, deformation may generate single-bed ramps, meaning ramps that cut across a single strong bed and die out in the weak bed above or below (Figure 18.23b). Displacement on wedge thrusts is generally less than the thickness of the bed.

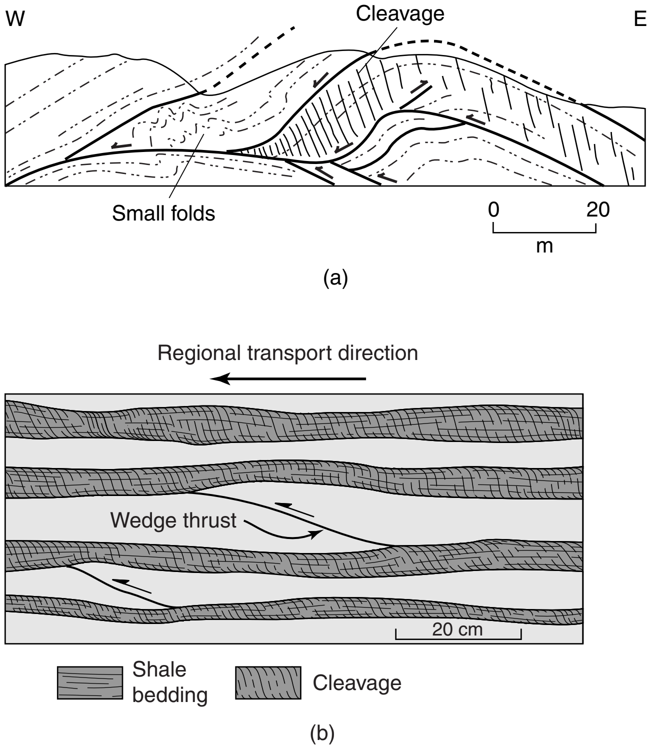


FIGURE 18.23 (a) Photo of a wedge thrust exposed on New World Island (Newfoundland). (b) Wedge thrusts forming in rigid limestone beds sandwiched between weak shale layers. In the shale layers, a cleavage has developed. The cleavage tips toward the hinterland.

4. *Grain-scale strain.* Rocks within thrust sheets locally undergo distortion at the scale of individual grains as a consequence of pressure solution between grains and/or plastic deformation within grains (e.g., by creating twins in calcite grains and deformation bands in quartz; see Chapter 9). The resulting strain ranges from a few percent to as much as 50%. Detection of grain-scale strain requires application of methods such as Fry analysis (see Chapter 4).
5. *Joints.* Fold-thrust belts typically contain systematic and non-systematic joint sets. The nonsystematic joints may develop in regions of tight folding, or adjacent to faults. Simplistically, we can divide systematic joints into two categories: strike-parallel joints and cross-strike joints. Chapter 7 provides a discussion of the origin of these joints.

18.6 FOLD-THRUST BELTS IN MAP VIEW

Faults are not surfaces with infinite dimensions; that is, the map trace of a fault must terminate along its length, either because the fault merges with or is truncated by

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[13] Some authors refer to the traces of folds, faults, and fabrics as the **trend lines** of an orogenic belt. Further, some authors refer to the pattern of trend lines as the **structural grain** of the belt.

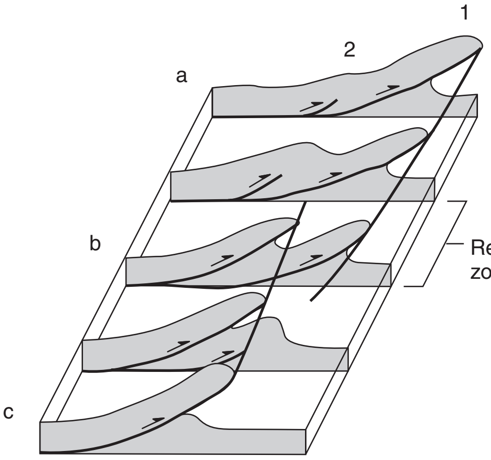

FIGURE 18.24 Concept of displacement transfer between thrusts in a simple relay zone. (a) The majority of displacement occurs on fault 1 and little occurs on fault 2. (b) Displacement is equally partitioned between faults 1 and 2. (c) The majority of displacement occurs on fault 2, little on fault 1.

another fault, or because the magnitude of displacement across the fault decreases progressively along strike until, at a tip line, displacement is zero. Beyond the tip line of a fault, shortening may be accommodated instead by folding or by slip on neighboring faults. For example, as shown in Figure 18.24, a decrease in slip on fault 1 is matched by an increase in slip on fault 2. Slip transfers from one fault to another at a **relay zone** or **transfer zone,** much like a baton transfers from one racer to the next in a relay race (Figure 18.24).

Commonly, the map trace of a major thrust or thrust system is convex toward the foreland (Figure 18.25a). According to the "bow-and-arrow rule" of thrusting, if you connect the termination points of the bowed fault trace with a straight reference line, the regional transport direction on the fault lies in a direction roughly perpendicular to this line. Further, the displacement magnitude on the fault is largest where the distance between the reference line and the fault trace is greatest (Figure 18.25b).

As illustrated by the Appalachians of eastern North America, regional map traces of fold-thrust belts typically are sinuous (Figure 18.7a). A place where the belt bulges into the foreland is a **salient,** whereas a place where the belt has not propagated so far into the foreland is a **recess.** The curved traces of structures in fold-thrust belts stand out in satellite images (Figure 18.26).

The origin of fold-thrust belt curvature remains controversial. Curves may develop (Figure 18.27a–f)

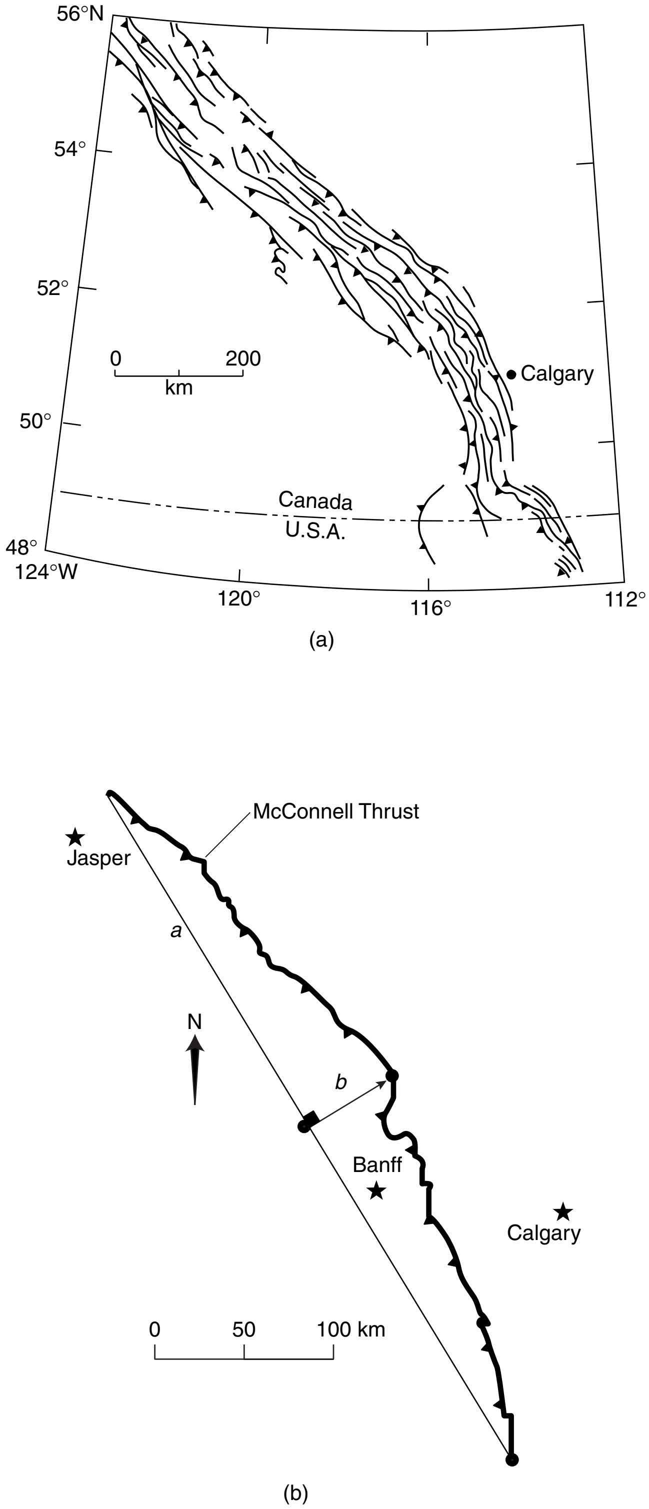

FIGURE 18.25 (a) Schematic map showing the traces of thrust faults in the southern Canadian Rockies. (b) The "bow-and-arrow rule," as applied to the McConnell Thrust, Alberta (Canada). For most foreland thrust belts, the ratio *b/a* is roughly 0.07–0.12.

FIGURE 18.26 Landsat satellite image of the curved Monterrey Salient (Mexico). The city of Monterrey is on the northern edge of the image. Long dimension of the photo is approximately 100 km.

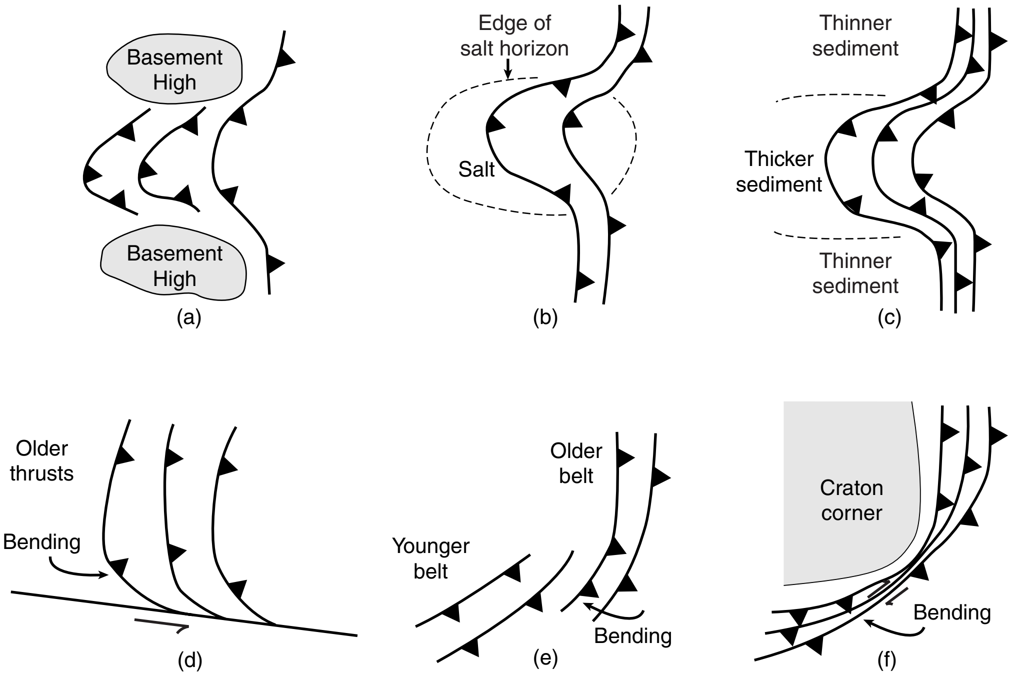

FIGURE 18.27 Map-view sketches illustrating possible processes leading to the formation of curved fold-thrust belts. (a) Interaction with basement highs in the foreland. New thrusts forming between the two highs originate with curved traces. (b) Lateral pinch-out of a stratigraphic glide horizon. The thrusts propagate further to the foreland over the weak salt horizon and the thrusts originate with curved traces. (c) Lateral variations in stratigraphic thickness. The thrusts propagate further to the foreland over the region where predepositional strata are thicker, and the thrusts originate with curved traces. (d) Interaction with a strike-slip fault. Motion on the strike-slip fault oroclinally bends the thrust traces. (e) Overprinting of two nonparallel thrust belts. Development of the younger belt bends the traces of the older belt. (f) Impingement against an irregular cratonic margin. Thrust sheets bend as they wrap around the corner.

where: (1) the belt interacts with basement highs in the foreland, for the basement highs retard propagation of thrusts; (2) the strength of the detachment horizon beneath changes along strike, because thrusts propagate further to the foreland where the detachment is weaker; (3) there are lateral variations in the thickness of the stratigraphic sequence being deformed, for thrust belts are wider where they involve thicker stratigraphic successions; (4) preexisting thrusts are bent by interaction with a strike-slip fault that cuts across the belt; (5) a second phase of thrusting overprints a preexisting belt at a high angle to the original belt; and (6) a fold-thrust belt impinges on the corner of a rigid craton.

Note that in some cases, the map-view curvature of fold-thrust belts develops when the thrusts initiate, so that right from the start the thrust has a curved trace. But in other cases, a preexisting straight thrust belt undergoes bending in map view during a second phase of deformation. A map-view curve that forms by the bending of a preexisting straight belt, such that the arms of the curve rotate around a vertical axis, is called an **orocline.**

18.7 BALANCED CROSS SECTIONS

In this chapter, we have presented several cross sections depicting the subsurface geometry of fold-thrust belts, right down to the basal detachment, even where these depths are not exposed. Perhaps you've asked yourself the fundamental question, "How do people draw such cross sections?" and "How reliable are they?" Well, to begin with, it is important to remember that a cross section is just an *interpretation* of the subsurface geology, and nothing more. We do not have access to outcrops several kilometers below the ground surface to let us see exactly where formation contacts, faults, and cutoffs are positioned. Cross-section interpretations are constrained by projecting surface geology into the subsurface, by interpreting seismic-reflection profiles, and by interpreting well data. Such data rarely provide a complete picture of subsurface geology, so we always must extrapolate when making cross sections. However, geologists have established a set of tests that permit us to evaluate cross sections to determine if the sections at least have a good chance of being correct. A cross section that passes these tests is said to be a **balanced cross section.** A balanced cross section has a reasonable chance of being correct, though we cannot guarantee it, whereas an unbalanced cross section is probably wrong (unless a good explanation can be provided for why the section does not balance).

Let's now look at four fundamental tests that help determine whether a cross section is balanced. Of note, these observations only apply when deformation does not result in movement in or out of the cross-section plane. Thus, they only apply to cross sections that have been drawn parallel to the transport direction on faults.

1. *The deformed-state cross section must be admissible.* Structures in the deformed-state cross section (i.e., the cross section depicting the way structures look today, after deformation) must resemble real structures that geologists have observed in outcrop or seismic profiles. For example, ramps should cut upsection, not downsection, unless they are out-of-sequence faults. We call cross sections that pass this test **admissible sections.** Figure 18.28a provides an admissible deformed-state cross section of the Lewis thrust sheet in Canada.
2. *Restoration of the cross section must yield reasonable geometries.* A **restored cross section** (Figure 18.28b) represents the predeformation configuration of strata and the predeformational location of faults in the region. "Restoration" of a cross section involves returning beds to horizontal by removing the effect of folding, by returning rocks to their original locations by removing the displacement on faults, and by undistorting thrust sheets by removing the effect of mesoscopic-scale and microscopic-scale deformation. The restored section must depict realistic-looking structures. For example, if a restored fault trace zigs and zags upsection, then there's likely something wrong with the section.
3. *The cross section must be "area balanced."* The area of rock shown on the restored cross section must equal the area shown on the deformed-state cross section unless pressure solution causes volume-loss strain. Cross sections that meet this criteria are **area balanced.** If volume-loss strain developed during deformation, this must be taken into account when restoring the section.
4. *The cross section must be kinematically reasonable.* It should be possible to create the deformed-state cross section from the restored cross section in a *kinematically reasonable way.* This means that you should be able to draw a series of cross sections depicting stages in the evolution of originally horizontal beds into the faulted and folded beds of the deformed-state cross section.

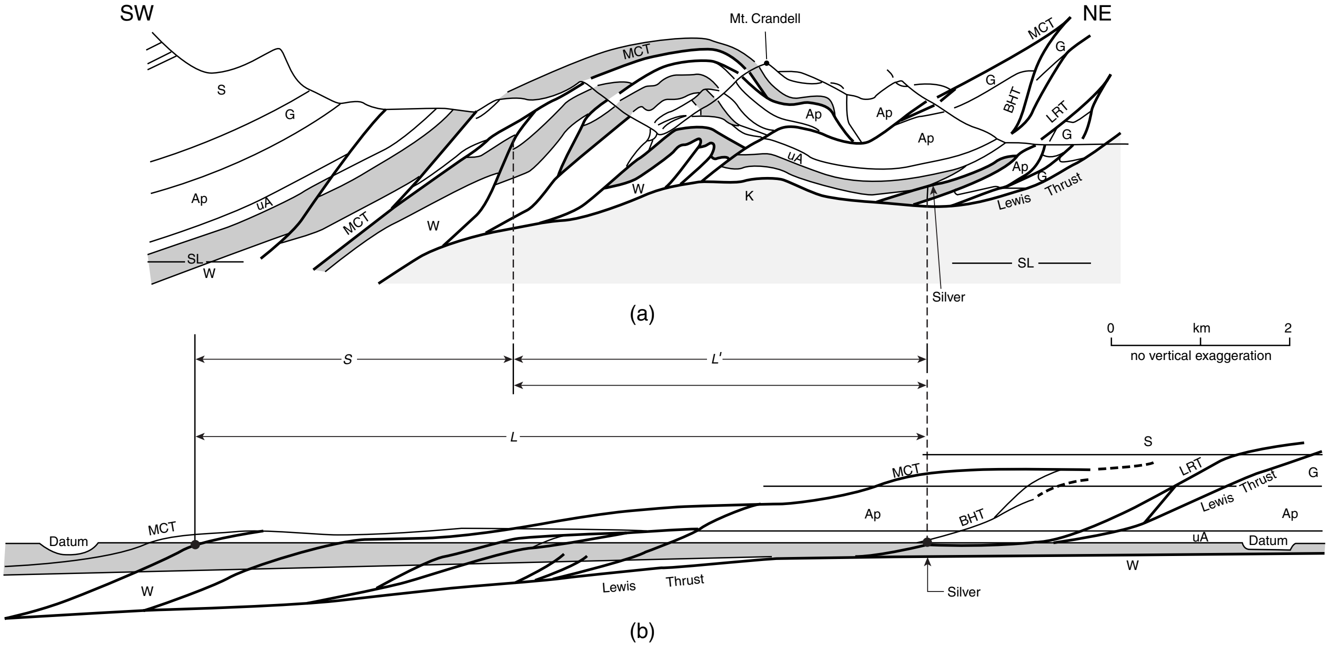

FIGURE 18.28 (a) Deformed-state cross section of a duplex within the Lewis Thrust Sheet, Waterton (Canada). No vertical exaggeration. (b) Restored version of the cross section. Hanging-wall strata are composed of the Precambrian Belt Supergroup (W = Waterton; shaded = lower Altyn; uA = upper Altyn; Ap = Appekunny; G = Grinnell; S = Siyeh; SL = sea level; MCT = McConnell thrust) that overlie footwall Cretaceous siliciclastics (K) across the Lewis Thrust. Shortening *(S)* is determined by comparison of the deformed and restored cross sections using the equation: $S = L - L'$. Note that this previously published cross section and restoration does not exactly balance in the foreland; to see why, try to match slices in the deformed and restored cross sections. Datuum is taken as the top of the shaded reference horizon.

The first two criteria in the above list ensure that the section doesn't depict impossible structural geometries. The third criterion ensures that the configuration of structures shown on the deformed-state section does not imply that undocumented volume change occurred during deformation. The fourth criterion emphasizes that you do not really understand the geometry of a complex structure until you can demonstrate how the structure formed from undeformed rock. "Balancing" a cross section involves the following steps. First, you carefully examine the deformed-state section for admissibility. Then, you construct a restored section and check it for admissibility and area balance. Finally, you think through a scenario that can explain the evolution of the deformed-state cross section from initially horizontal beds.

It is beyond the scope of this chapter to provide detailed guidelines for balancing cross sections; most structural geology laboratory manuals offer step-by-step instructions and exercises. But we do point out that, in some cases, you can use **quick-look techniques** to quickly scan a deformed-state cross section and determine if it has the potential to be balanced. To apply these techniques, first identify ramps and flats in each part of the cross section, and count them (Figure 18.29). Are there the same number of ramps and flats in the hanging wall as in the footwall? There should be, because in an admissible, restored cross section, the hanging wall fits over the footwall with no gaps or overlaps. Now, paying particularly close attention to the ramps, check to see if the same beds are truncated in the hanging wall as in the footwall. They must be, because the hanging-wall beds were originally adjacent to the footwall beds. These two simple tests will highlight the majority of common cross-section errors that lead to construction of unbalanced cross sections.

Let's apply the quick-look technique to the example depicted in Figure 18.28. First, note that the sliver of lower Altyn Formation (shaded), between faults *BHT* and *LRT* in the deformed section, has a short hanging-wall ramp and a long hanging-wall flat (Figure 18.28a). In the restored section (Figure 18.28b), however, this sliver only has a hanging-wall ramp. Further, in the restored section, the area of the sliver is much smaller than in the deformed-state section. Thus,

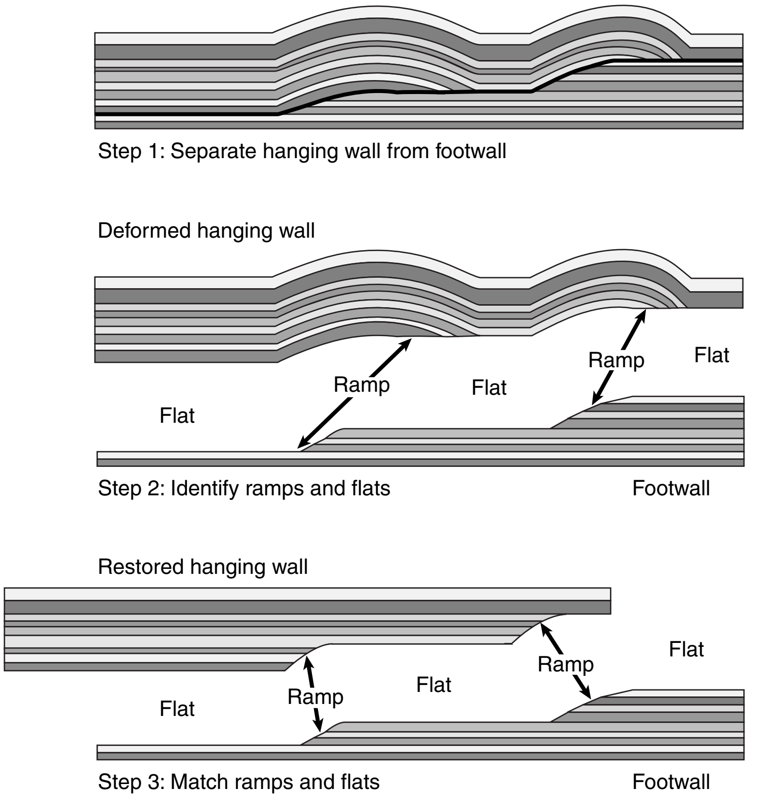

FIGURE 18.29 Diagram illustrating quick-look technique for checking a cross section for potential problems. The key is to recognize ramps and flats in the deformed-state section and realize that hanging-wall flats and ramps must exactly match footwall ramps and flats in number and in stratigraphic composition.

this part of the deformed-state cross section cannot be balanced. There are other mismatches on this section, as well, so the geologist constructing this section should check to determine if there has been movement out of the plane of section, if there is a drafting error, or if there is an alternative interpretation that can be balanced.

Before getting too carried away with the value of cross-section balancing, keep in mind that not all cross sections have to balance. Two-dimensional balancing techniques cannot deal with cross sections over lateral or oblique ramps, or across strike-slip faults. In such settings, movement takes place in or out of the plane of the section, and thus by definition, area balance in the cross-sectional plane is impossible. Similarly, in regions where deformation has occurred by flow of rock, units may be sheared and isoclinally folded, or may flow into or out of the section plane, to an extent that balance is again impossible. Even in low-grade rocks, pressure solution may cause significant volume change that may make area balance a challenge.

We conclude our discussion of balancing by once again pointing out that checking a cross section for balance does not automatically ensure a "correct" interpretation or mean that the interpretation is unique (i.e., there may be other balanced interpretations that fit the data). Balancing procedures are simply meant to focus your attention on potentially problematic areas in the cross section that require geologic explanation and/or reinterpretation.

18.8 MECHANICS OF FOLD-THRUST BELTS

Now that we have a feel for the geometry of structures that occur in a fold-thrust belt, we can explore the mechanisms of fold-thrust belt development. The formation of fold-thrust belts has been particularly perplexing to geologists because the movement of large thrust sheets—tens of kilometers wide as measured in the transport direction, but less than a few kilometers thick—at first seems like a paradox. Picture a thrust sheet, in cross section, to be a rectangle resting on a surface (Figure 18.30a). If you assume a reasonable value for frictional resistance (σ_f) to sliding on the underlying detachment, assuming the detachment is dry, then the stress (σ_1) necessary to push the sheet over a horizontal surface or up a gentle incline greatly exceeds the failure strength of the intact rock comprising the thrust sheet (see Chapter 6). Thus, you would expect the hinterland end of the thrust sheet to crush or buckle before the sheet as a whole would move. Yet clearly, large thrust sheets do exist. So how do they move, and how do belts composed of many large thrust sheets develop? It took several decades for geologists to develop models that address these questions.

Geologists first attacked the issue of how to overcome the friction between solid surfaces that presumably provided the resistance for thrust-sheet movement. In the late 1950s researchers[14] realized that the force required to move a thrust sheet can be greatly diminished if hydrostatic pressure in the detachment zone increases to values approaching lithostatic loads (Figure 18.30b). "Hydrostatic pressure" refers to the pressure in the water that fills pores and cracks in rock, while "lithostatic pressure" refers to the pressure within solid grains generated where grains are in direct contact. A pore resembles a tiny balloon between

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[14] M. King Hubbert and William Rubey provided some of the key contributions to this idea.

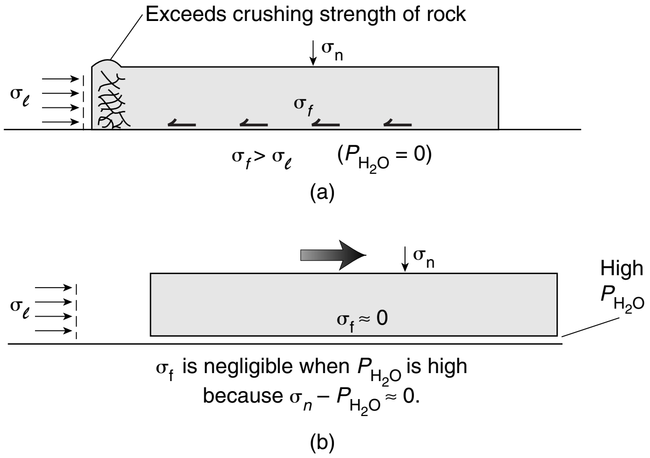

FIGURE 18.30 (a) When pushed from the rear, a rectangular thrust sheet sliding on dry rock would be crushed before overcoming frictional resistance. (b) If there is high fluid pressure at the basal detachment, the effective stress decreases and allows the thrust sheet to move under very small applied load. σ_l is the stress resulting from horizontal loading, σ_f is frictional resistance, σ_l represents the boundary load at the end of the thrust sheet, and P_{H_2O} represents the pore pressure. Other terms are defined in the text.

grains—if you force water into the pore, the water pushes outward, just like the air in a balloon pushes outward if you blow up the balloon. In effect, the fluid pressure in a detachment zone "lifts up" the thrust sheet so that it can glide over the detachment.

By looking once again at the Mohr-Coulomb criterion for failure (Chapter 8), we can gain further insight into the role of fluid pressure in thrust-sheet movement. For sliding to occur, the shear stress (σ_s) applied to the fault surface must exceed frictional resistance (σ_f). When rock is dry, the equation relating the shear stress (σ_s) necessary to cause sliding on a detachment surface to the normal stress (σ_n) squeezing the two sides of the detachment together can be written

$$\sigma_s = C + \mu\sigma_n \qquad \text{Eq. 18.1}$$

where C is the cohesion and μ is the coefficient of internal friction. This equation shows that as the normal stress (representing the vertical load due to the weight of the thrust sheet) increases, the shear stress needed to cause movement on the detachment increases. If, however, the rock in the detachment horizon is wet, water creates pore pressure. The pore pressure pushes up the load, and thus counteracts σ_n. Thus, the frictional resistance to sliding depends instead on the **effective normal stress** (σ_n*) across the surface. We define σ_n* by the equation

$$\sigma_n^* = \sigma_n - P_{H_2O} \qquad \text{Eq. 18.2}$$

where P_{H_2O} is the fluid pressure. Substituting effective stress (σ_n*) back into the Mohr-Coulomb criterion yields

$$\sigma_s = C + \mu\sigma_n^* \qquad \text{Eq. 18.3}$$

Equation 18.2 indicates that as fluid pressure increases, the effective normal stress across the detachment decreases, because fluid in the detachment zone partially supports the weight of the thrust sheet. Equation 18.3 shows that as the effective normal stress decreases, the shear stress needed to cause sliding decreases. Therefore, the thrust sheet can be moved easily when fluid pressure is high, even when the boundary load at the hinterland edge of the sheet is significantly lower than the failure strength of rock comprising the sheet.

But what initiates thrust motion? Initially geologists assumed that thrust sheets slid toward the foreland in response to gravity when the basal detachment dipped toward the foreland, and thus that thrust sheets moved like slumps on a hillslope. Such **gravity sliding** models became very popular as a cause for thrusting, and in the 1960s, most structural geologists envisioned that development of fold-thrust belts occurred as thrust sheets glided down an incline created by uplift of the hinterland during orogeny (Figure 18.31a and b). However, petroleum exploration of many fold-thrust belts provided seismic-reflection data showing that basal detachments beneath almost all classic fold-thrust belts dip toward the hinterland, not the foreland! To account for this contradiction, some structural geologists suggested that fold-thrust belt formation was a consequence of **gravity spreading,** meaning that fold-thrust belts form when the thickened crust of an orogen "collapses" and spreads laterally under its own weight, much like a continental ice sheet spreads away from the region where snow accumulates (Figure 18.31c). The main difference between the gravity-spreading model and the gravity-sliding model is that the former implies that the direction of dip of the topographic surface of the fold-thrust belt, not the dip of the basal detachment, drives thrust movement.

The next step in formulating an understanding of how fold-thrust belts develop came in the 1970s, when researchers began to study laboratory models that simulated the development of the belts. Models involving the formation of a sand wedge building in front of a plow proved to be particularly informative (Figure 18.32). That's because sand is a **Coulomb material,** meaning an aggregate composed of grains that can frictionally slide past one another, and at the scale of a mountain range, rock of the upper crust behaves

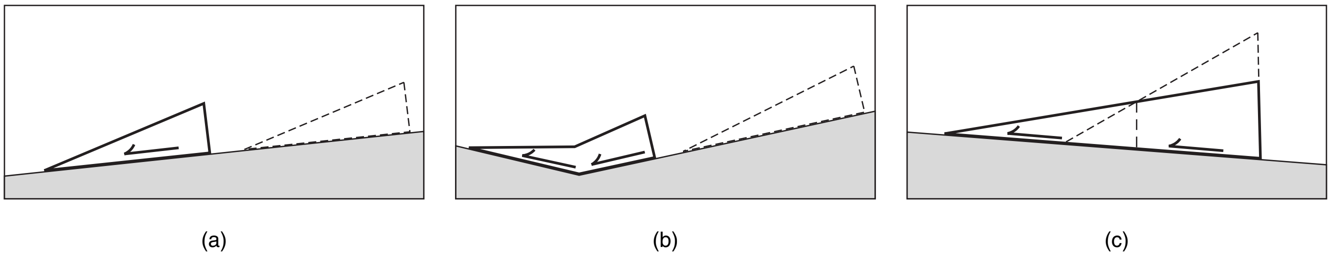
(a) (b) (c)

FIGURE 18.31 (a) The concept of gravity sliding. Here a block slides down a foreland-tilted slope. The dashed lines show the original position. (b) Gravity sliding partly downslope and partly upslope. The dashed lines show the original position. (c) The concept of gravity spreading. Before spreading, the wedge had the shape indicated by the dashed line.

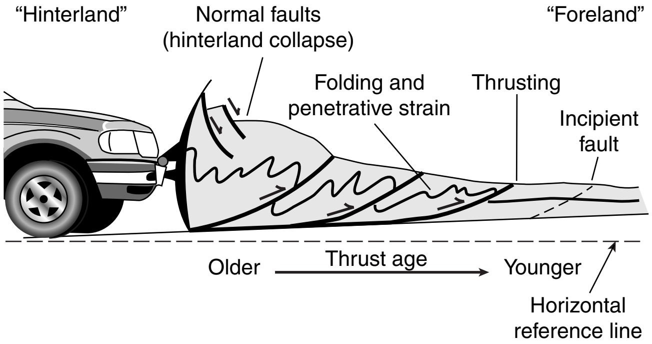

FIGURE 18.32 Snowplow analogy for fold-thrust belt development. The wedge of snow widens with continued shortening; younger thrusts generally initiate in a hinterland to foreland progression. While new thrusts are adding material at the toe of the wedge, the hinterland portions are developing penetrative strain, normal faults, and slump features.

essentially like a Coulomb material. In typical sand-wedge models (or, more generally, Coulomb-wedge models), the plow pushes into a sand layer underlain by a hard material. The boundary between the sand and the underlying hard material represents the basal detachment. Note that the wedge depicted in Figure 18.32 has a foreland-dipping topographic surface and a hinterland-dipping basal detachment—this shape is similar to the overall cross-sectional shape of a fold-thrust belt.

Two sources of stress drive the development of a Coulomb wedge. One source is a result of the displacement of the plow blade toward the foreland. This stress is called a **horizontal boundary load.** Another source is the result of the gravitational potential energy that develops when a foreland-dipping topographic slope develops (i.e., when the hinterland portion of the wedge rises and becomes higher than the foreland portion). Note that the gravitational potential energy, due to the elevation of the hinterland, creates both vertical and horizontal stresses. In Figure 18.33a, we represent these stresses as follows: σ_{bs} is the horizontal boundary load caused by movement of the backstop toward the foreland; σ_{gv} is the vertical component of stress caused by gravity; σ_{gh} is the horizontal component of stress caused by gravity.

Let's look more closely at the evolution of a Coulomb wedge (e.g., a sand wedge). As the backstop moves toward the foreland, the wedge deforms internally (by forming folds, faults, and grain-scale distortion) and, as a consequence, its surface slope increases. When the wedge reaches a certain **critical taper angle,** ϕ_c (defined as the surface slope angle, α_1, plus the detachment dip, β), the wedge as a whole slides toward the foreland along the weak detachment (Figure 18.33a). Slip occurs on the detachment because the coefficient of sliding friction on the detachment is less than the coefficient of internal friction in the wedge. If the taper angle becomes too large (Figure 18.33b), processes take place within the wedge to cause the surface slope of the wedge to decrease (Figure 18.33c). Several processes can cause the slope angle to decrease, including the addition of new thrust slices at the toe (a process called offscraping), the erosion of the higher portions of the wedge, or the development of extensional faulting (extensional collapse) within the wedge. If these processes continue until the taper angle become less than the critical taper angle (Figure 18.33d), sliding of the wedge stops, and deformation within the wedge occurs once again to thicken the wedge internally and increase the surface slope (Figure 18.33e). This internal thickening can involve reactivation of thrusts, formation of out-of-sequence faults, formation of duplexes at the base of the wedge (a process also called underplating), or formation of penetrative strain and folding within thrust sheets. Internal thickening increases the topographic slope angle until the wedge achieves the critical taper angle again. Then, the wedge again starts sliding toward the foreland and new thrusts again form at the toe.

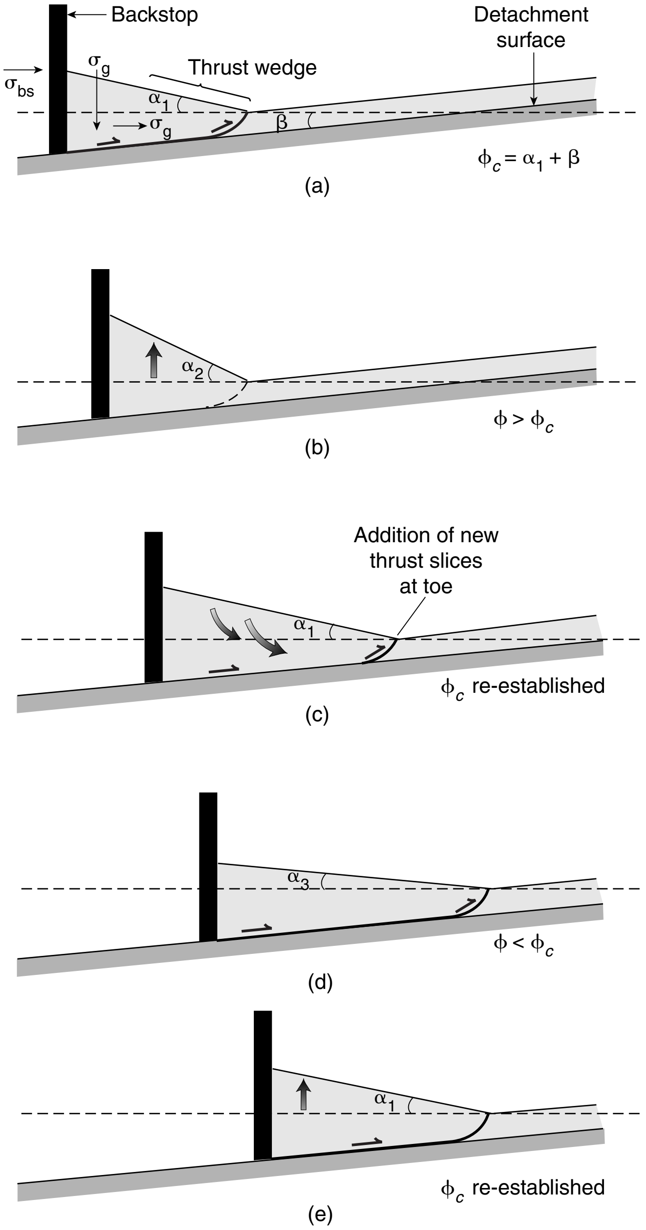

FIGURE 18.33 The critical taper theory of fold-thrust belt mechanics. The critical taper (ϕ_c) is defined as the sum of the surface slope angle (α_1) and the detachment slope angle (β). (a) Stress acting on a wedge is partly a horizontal boundary load caused by the backstop (σ_{bs}) and is partly caused by gravity (σ_g). (b) If the backstop moves, the wedge thickens, so the surface slope increases, and the taper (ϕ) eventually exceeds ϕ_c. (c) The wedge slides toward the foreland and new material is added to the toe, and extension of the wedge occurs so that surface slope decreases. (d, e) If the surface slope becomes too small, thrusting at the toe stops, and the wedge thickens by penetrative strain or out-of-sequence thrusting.

The observed critical taper of a given fold thrust belt depends on the material strength of the wedge, the resistance to sliding across the basal detachment, and the ratio of fluid pressure to overburden pressure both in the wedge and across the detachment. If the effective strength of the wedge is increased (either by increasing rock strength or by decreasing fluid pressure), then the critical taper angle decreases. If the resistance to sliding on the basal detachment is increased (either by increasing the coefficient of sliding friction or by decreasing the fluid pressure), then the critical-taper angle increases. Fold-thrust belts whose basal detachments lie in salt, a very weak lithology, have a critical taper angle as low as 1° to 2°, whereas fold-thrust belts that have detachments in stronger rocks may have critical taper angles as high as 8° to 10°.

Overall, we see that in the Coulomb-wedge model of fold-thrust belt development, now known formally as **critical taper theory,** the belt evolves and grows by maintaining a dynamic equilibrium between (1) addition of new material at the toe of the wedge, which decreases the surface slope, (2) internal deformation within the wedge, which increases the surface slope, and (3) thinning of the wedge due to extensional deformation (i.e., hinterland collapse) and/or erosion of the wedge.

Can critical taper theory explain the break-forward sequence of thrusting in a fold-thrust belt? Yes. Imagine a plow at the end of a snow-covered concrete driveway (the "basement") that slopes gently toward the street (Figure 18.32). As soon as the plow begins to move, a wedge-shaped pile of snow forms in front of the blade, and the surface of this pile slopes toward the foreland. A detachment forms beneath the pile, and a thrust develops at the front of the pile. A few meters in front of the plow, however, the snow remains totally unaffected. During the next increment of movement, the detachment propagates farther beneath the snow and above the concrete "basement" toward the foreland, and a new thrust develops. The thrust system now consists of two imbricate thrusts cutting upward from the basal detachment, with the imbricates forming in a break-forward sequence. As the process continues, an imbricate fan of thrusts develops, with the youngest thrust furthest to the foreland. In other words, Coulomb wedge models demonstrate that thrusting occurs in a break-forward sequence. Note that, because Coulomb materials are not very strong, the wedge cannot thicken indefinitely. Eventually, the hinterland of the wedge collapses under its own weight by slumping or by formation of normal faults within it. This hinterland collapse results in a net thinning of the thicker part of the wedge, thereby maintaining a dynamic

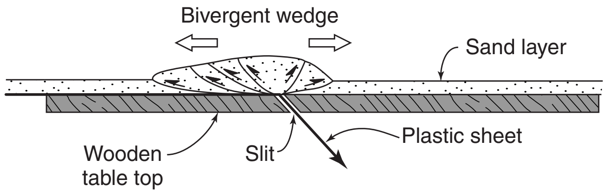

FIGURE 18.34 Simplified cross section of a sandbox model illustrating development of a bivergent thrust wedge. The sand was laid down on a mylar (plastic) sheet, and the sheet is pulled through a slit in the wooden table beneath it. Note that the wedge is not symmetrical.

balance between wedge thickening and the proper surface slope for wedge translation.

In recent years, geologists have modified Coulomb-wedge model design to make the models more realistic. In the newer models, there is no rigid backstop. Rather, sand is placed on a thin plastic sheet, and the sheet is then pulled through a slit at the bottom of the sand box. Such a model configuration resembles what happens where one plate slides beneath another at a convergent or collisional boundary. Note that the sand itself serves as the backstop. As motion progresses, the sand on the plastic sheet (i.e., the sand on the downgoing or underthrust plate) pushes against the stationary sand on the overriding plate. As a result, a **bivergent wedge** of sand develops (Figure 18.34). This consists of two thrust belts on opposite sides of the slit, each of which verges away from the slit. The "forewedge," which verges toward the interior of the downgoing plate, tends to be wider than the "retrowedge," which verges toward the interior of the overriding plate. The bivergent-wedge model explains the geometry of thrusting at some accretionary prisms, in which thrusting toward the sea occurs at the toe of the prism, while thrusting toward the arc occurs at the hinterland portion of the wedge. This model also provides an analog for the gross geometry of a collisional orogen, for in many such orogens, thrust belts form on both sides of the metamorphic hinterland.

18.9 CLOSING REMARKS

Fold-thrust belts are inherently fascinating geologic terranes. They contain all the components that make for a good scientific puzzle—intriguingly complex features and potentially quantifiable relationships. In addition, they yield beautiful mountain ranges that are exciting settings for field work, and which may contain valuable resources. No wonder fold-thrust belts have been the focus of such intense research for so long, and will undoubtedly challenge the talents of geologists for years to come.

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