CHAPTER TWENTY-TWO

Western Hemisphere

22.1 THE NORTH AMERICAN CORDILLERA—An essay by Elizabeth L. Miller[1]

22.1.1 Introduction

The broad mountainous region of western North America is known as the **"Cordillera,"**[2] an orogenic belt that extends from South America (the Andean Cordillera) into Canada (Canadian Cordillera) and Alaska (Figure 22.1.1). The youthful topography of this impressive mountain belt is closely related to ongoing crustal deformation, as indicated by the distribution of seismicity across the width of the orogenic belt (Figure 22.1). The present plate tectonic setting and the dominant style of deformation vary along strike of the orogen: folding and faulting above an active subduction zone in the Pacific Northwest of the United States and Alaska; strike-slip or transform motion along the Queen Charlotte Fault (Canada) and San Andreas Fault (California); and extension and rifting in the Basin and Range Province of the western United States and Mexico's Gulf of California. Variations in structural style are also apparent across the orogen; for example, crustal shortening and strike-slip faulting in coastal California are concurrent with crustal extension and basaltic volcanism in the adjacent, and inboard, Basin and Range Province. This diversity in plate tectonic setting and structural style of the Cordillera along and across strike, likely characterized the past history of the orogenic belt, which has been shaped primarily by Pacific–North America plate interactions. The continuity of such interactions since the Late Precambrian makes it the longest-lived orogenic belt known on Earth.

The Cordillera provides an excellent natural laboratory for studying the evolution of a long-lived active margin and the effects of subduction and plate boundary processes on the evolution of continental crust. However, the exact nature of the relationship between plate motions and continental deformation, whether mountain building or crustal extension, remains a complex question for the following reasons. The theory of plate tectonics treats the Earth's lithosphere as a series of rigid plates that move with respect to one another along relatively discrete boundaries. This simplification applies well to oceanic lithosphere, which is dense and strong and thus capable of transmitting stresses across great distances without undergoing significant internal deformation. However, it does not apply well to continents, whose more quartzo-feldspathic composition and greater radiogenic heat flow make them inherently weaker. Displacements or strain within continental crust can accumulate at plate tectonic rates (1–15 cm/y) within narrow zones of deformation, or can take place more slowly (millimeters to centimeters per year) across broad zones of distributed deformation. Thus continents can accumulate large strains, thickening over broad distances during crustal shortening, and thinning during extension. Evidence for these strain histories are at least partially preserved in the geologic record because the inherent buoyancy of continental material prevents it from being subducted into the mantle. Mantle-derived magmatism can lead to greater strain accumulation within continental crust by increasing temperatures and thus rheologically weakening the crust, allowing it to deform in a semicontinuous fashion. Because of these considerations regarding the thermal structure, composition, thickness, and rheology of continental crust, the response of the overriding continental plate to changes in subducting plate motion or to changes in the nature of plate interactions along a margin may be sluggish and may vary with time and depth in a complex fashion.

Thus, orogenic belts like the Cordillera may be at best imperfect recorders of past plate motions. Our understanding of the link between plate tectonics and continental deformation is evolving as more detailed geologic and geochronologic studies are carried out,

[1]Stanford University, Stanford, CA.
[2]Spanish for mountain or mountain chain.

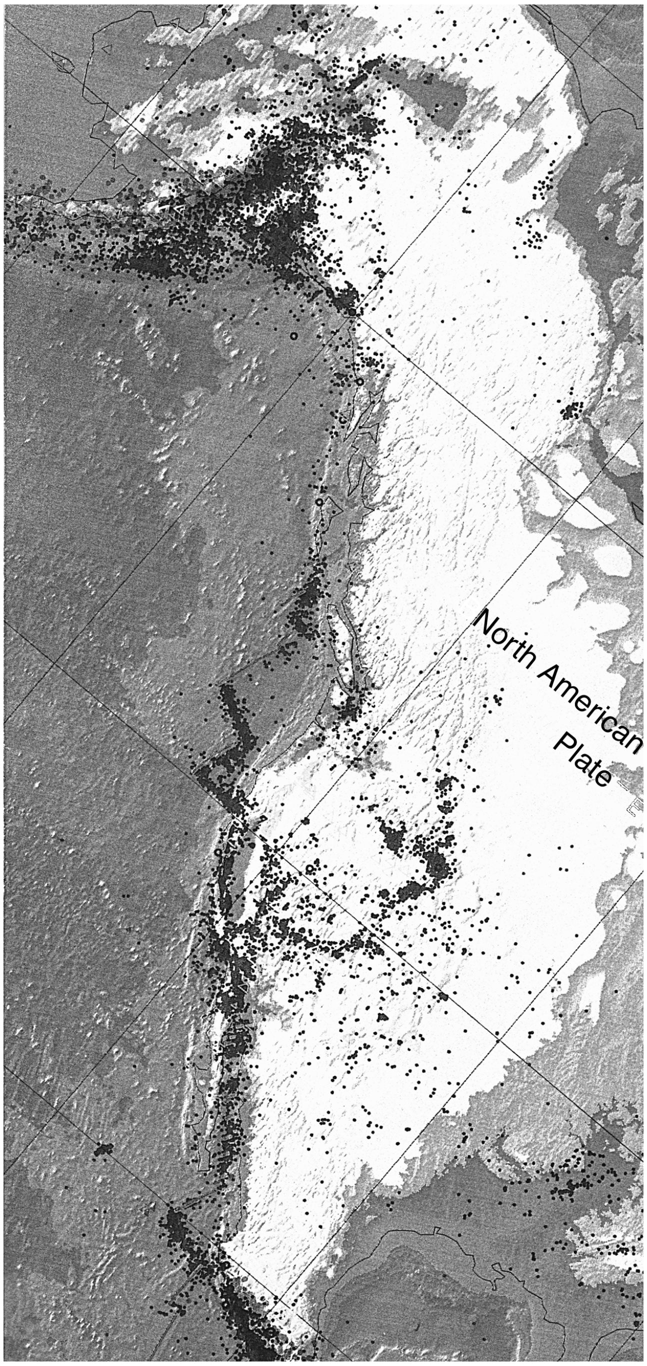


FIGURE 22.1.1 Digital topography, plate tectonic setting and seismicity map of the North American Cordillera.

providing quantitative information on the timescale of events and the rates of geologic processes, and increasing our ability to compare the timing of events from one part of the belt to the next. Geophysical and petrologic studies remain key tools that help us to understand how physical processes in the deeper crust and mantle are coupled to more easily studied deformation at shallower levels of the crust.

22.1.2 Precambrian and Paleozoic History

Studies of modern active plate margins have played an important role in interpreting the more fragmentary geologic record of analogous events in the North American Cordillera (Figure 22.1.2). Based on these comparisons, it appears that all plate tectonic styles and structural regimes known on earth, with the exception of continent-continent collision (like the Himalaya), played a part in the creation and evolution of the North American Cordillera. The initial formation of the Cordilleran margin dates back to the latest Precambrian. The Windemere Supergroup is a thick succession of shelf-facies clastic rocks deposited between about 730 Ma and 550 Ma, its facies and isopachs define the newly rifted margin of western North America after the breakup of the Rodinia supercontinent. The Windemere Supergroup forms the lower part of a 15-km thick, dominantly carbonate shelf succession whose isopachs and facies boundaries closely parallel the trend of the Cordillera and are remarkably similar along the length of the Cordillera from Mexico to Alaska. This shelf sequence (including the older and more localized Belt-Purcell Supergroup) is now spectacularly exposed in the eastern Cordilleran fold-and-thrust belt, whose overall geometry and structure are controlled by the facies and thickness of this succession.

The Paleozoic history of the Cordillera has generally been described as one of continued passive margin sedimentation and little active tectonism. However, embedded in the western Cordillera are a multitude of tectonic fragments of island arcs and backarc basin sequences that range in age from Cambrian to Triassic (Figure 22.1.2). The common conception that these represent a collage of far-traveled terranes exotic to the Cordillera ("suspect terranes") is being reevaluated as a result of numerous studies that indicate many of these sequences developed adjacent to, but offshore of, the western edge of the continental margin. Although these "exotic" fragments are likely displaced from their site of origin by rifting, thrusting, and strike-slip faulting, their presence nonetheless argues convincingly for a long history of subduction of paleo-Pacific crust beneath the western edge of the North American plate. Study of these accreted fragments suggests that during the Paleozoic, western North America looked much like the Southwest Pacific today, with its fringing arcs separated from the main Australasian continental shelves by backarc basins. During the Paleozoic, the North American shelf itself experienced episodes of regional subsidence and uplift, but no sig-

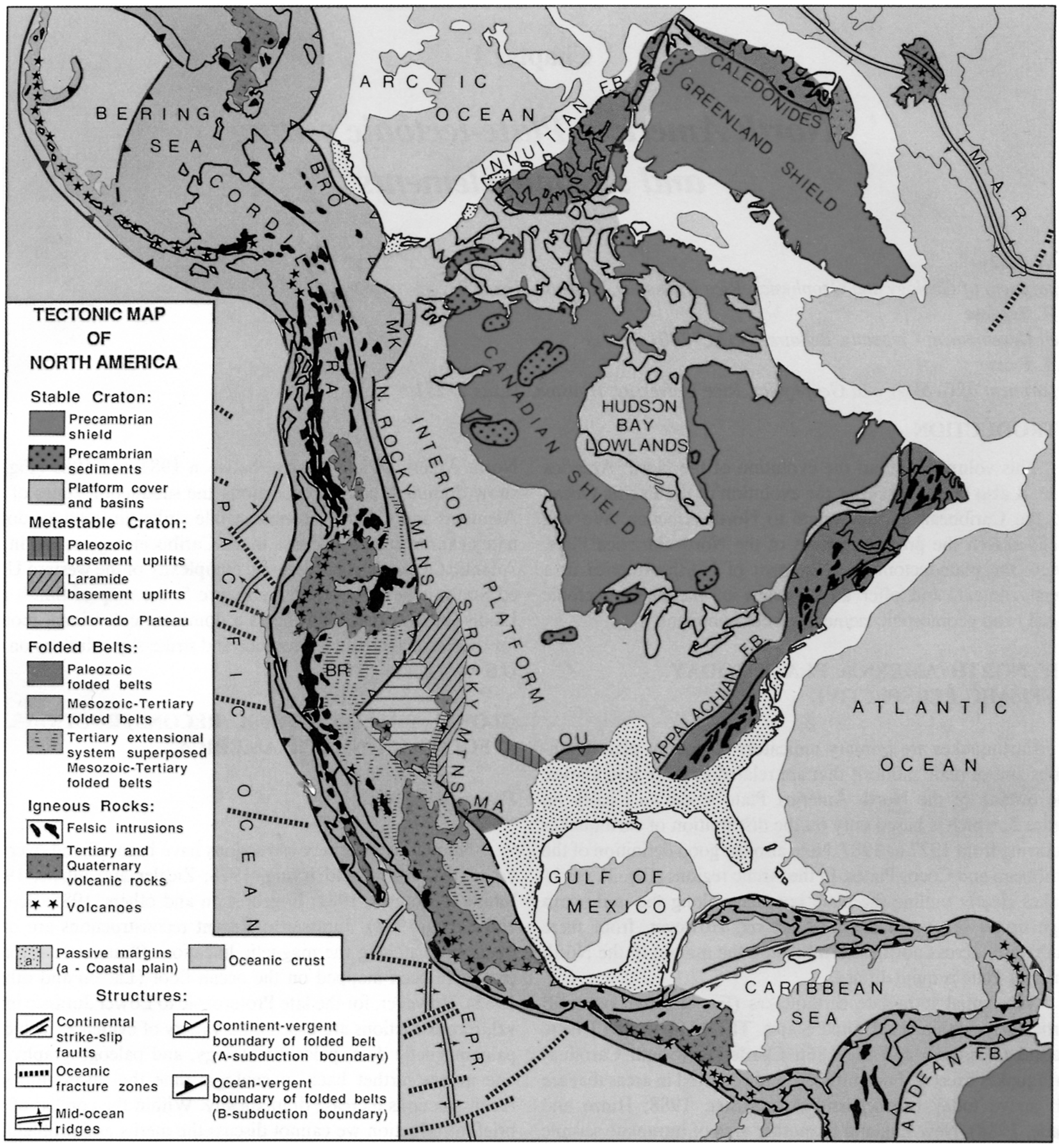


FIGURE 22.1.2 Simplified geologic/tectonic features map of North America. BR = Basin and Range Province; BRO = Brooks Range, MK = Mackenzie Mountains, MA = Marathon Uplift, OU = Ouachita Mountains, M.A.R. = Mid-Atlantic Ridge, E.P.R. = East Pacific Rise; F.B. = fold belt.

nificant deformation. Exceptions include deformation and intrusion of latest Proterozoic to Late Devonian granites along parts of the margin in Alaska and southern British Columbia, and the closure of deep-water, backarc basins by thrusting onto the shelf during the earliest Mississippian Antler Orogeny (Roberts Mountains Thrust) and during the Permo-Triassic Sonoma Orogeny (Golconda Thrust) in the western U.S. part of the Cordillera.

22.1.3 Mesozoic History

Magmatic belts related to eastward subduction beneath western North America are much better developed beginning in the Mesozoic (Figure 22.1.2). Arc magmatism of Triassic and Early Jurassic age (230–180Ma) is recorded by thick sequences of mafic to intermediate volcanic rock erupted in an island arc (Alaska, Canada, and the northern part of the U.S.) or continental arc (southwestern U.S.) setting. Tectonism accompanying subduction during this time-span was generally extensional in nature, leading to rifting and subsidence of parts of the arc and continental margin. The Middle to Late Jurassic brought a dramatic change in the nature of active tectonism along the entire length of the Cordilleran margin. This time-span is characterized by increased plutonism during the interval 180–150 Ma and many hundreds of kilometers of crustal shortening. This shortening closed intra-arc and backarc basins, accreted arc systems to the North

American continent, and fundamentally changed the paleogeography of the Cordillera from a southwest Pacific–style margin to an Andean-style margin, a tectonic framework that persisted throughout most of the latest Mesozoic and Cenozoic. The preferred explanation for this orogeny is that it is linked to rapid westward motion of North America with respect to a fixed hot-spot reference frame (Figure 22.1.3). This westward motion occurred as the North Atlantic began to open and, in effect, caused the western margin of the continent to collide with its own arc(s) and subduction zone(s) and then to deform internally. Deformation associated with Middle Jurassic orogenesis (sometimes referred to as the Columbian Orogeny) began first in the region of elevated heat flow within and behind the arc, and migrated eastward with time towards the continental interior.

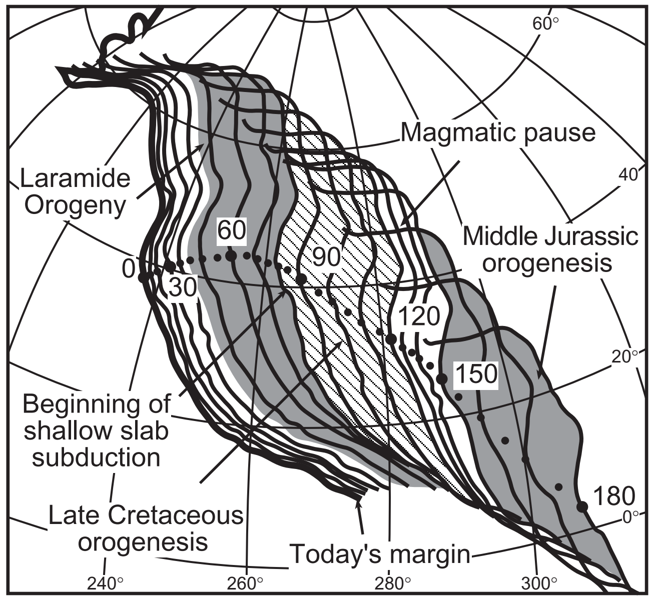
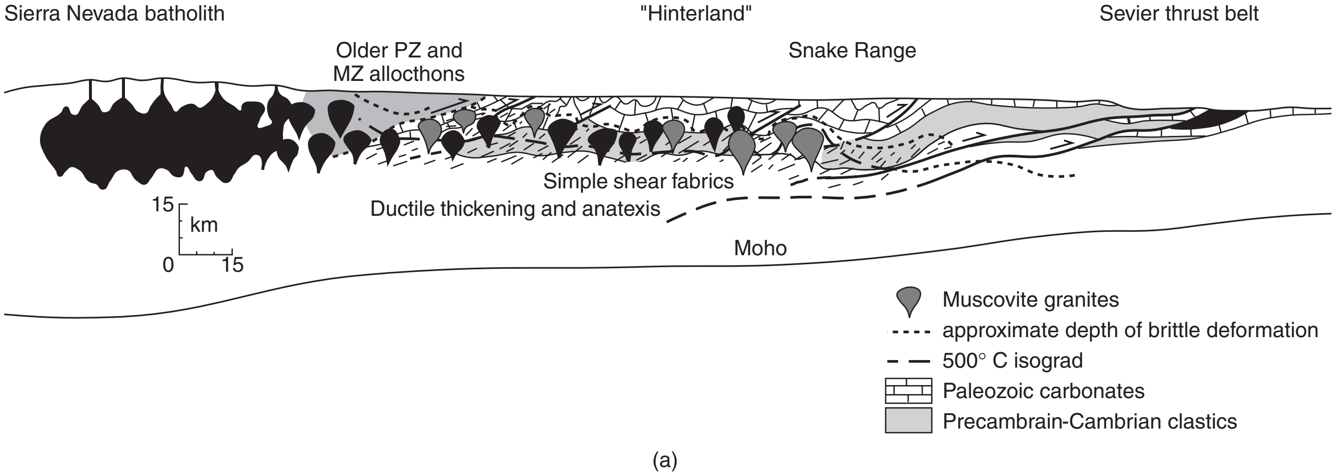


FIGURE 22.1.3 Correlation of deformational events and motion of the west coast of North America with respect to hot spots from the Late Jurassic to the Cenozoic.

In the Cretaceous, we have a better record of the nature of Pacific Plate motions with respect to North America (in large part from magnetic anomalies on the ocean floor) and it is possible to draw some inferences about the link between orogenic and magmatic events and the history of subduction beneath the active margin. There is a general lull in deformation and a lack of evidence for significant magmatism during the time interval 150–120 Ma, which possibly corresponds to a time of dominantly strike-slip motion along the Cordilleran margin (Figure 22.1.3). Particularly rapid rates of orthogonal subduction of the Farallon and Kula Plates occurred in the later part of the Cretaceous and Early Tertiary, resulting in the emplacement of major batholiths inboard of continental margin subduction zone complexes. Depending on the configuration of subducting oceanic plates, large components of margin-parallel strike-slip faulting are also implied. How much, where, and along what faults this motion was accomplished are still controversial questions. In the western United States, the emplacement of the composite Cretaceous Sierra Nevada batholith occurred (Figure 22.1.2). On the western (forearc) side of the batholith, depressed geotherms caused by the rapidly subducting slab led to high pressure–low temperature (blueschist) metamorphism within rocks now represented by the Franciscan Complex. The intervening Great Valley Basin underwent a similar history of "refrigeration" during rapid subduction. Sediments deposited in this basin were buried as deep as 10 km, but the section reached temperatures of only about 100°C, suggesting thermal gradients of 10°C/km or less. In contrast, heat flow in the arc and backarc region to the east was high and accompanied by major shortening; the latter migrated eastward with time and resulted in the well-known Sevier foreland fold-and-thrust belt (Figure 22.1.4). At any given latitude, there are typically a series of major thrusts that displace stratified Paleozoic-Mesozoic shelf sediments eastward, with a minimum total displacement of 100–200 km. Along most of its length, the eastern front of the thrust belt closely follows the transition from thin cratonic to thicker shelf sequences, indicating important stratigraphic influence on the structures produced. Deeper parts of the crust between the thrust belt and the magmatic arc were hot and mobile and underwent thickening by folding and ductile flow (Figure 22.1.4). Crustal thickening in turn precipitated crustal melting, now represented by a belt of unusual muscovite-bearing granites that lie just west of the main thrust belt (Figure 22.1.4). Because the orogen at this latitude was later reworked by Cenozoic extension and its crust thinned, the amount of crustal thickening during the Mesozoic is still controversial. Was the western United States, at the end of the Cretaceous like the Tibetan Plateau or Andean Altiplano, underlain by 70–80-km thick crust? Or was crustal thickening more modest as evidenced by the ~50 km thick crustal root beneath the unextended Canadian Cordillera today?

22.1.4 Cenozoic History

After the latest Cretaceous, the history of the western U.S. segment of the Cordillera differs substantially from its neighboring segments to the north and south, where subduction driven arc magmatism and crustal



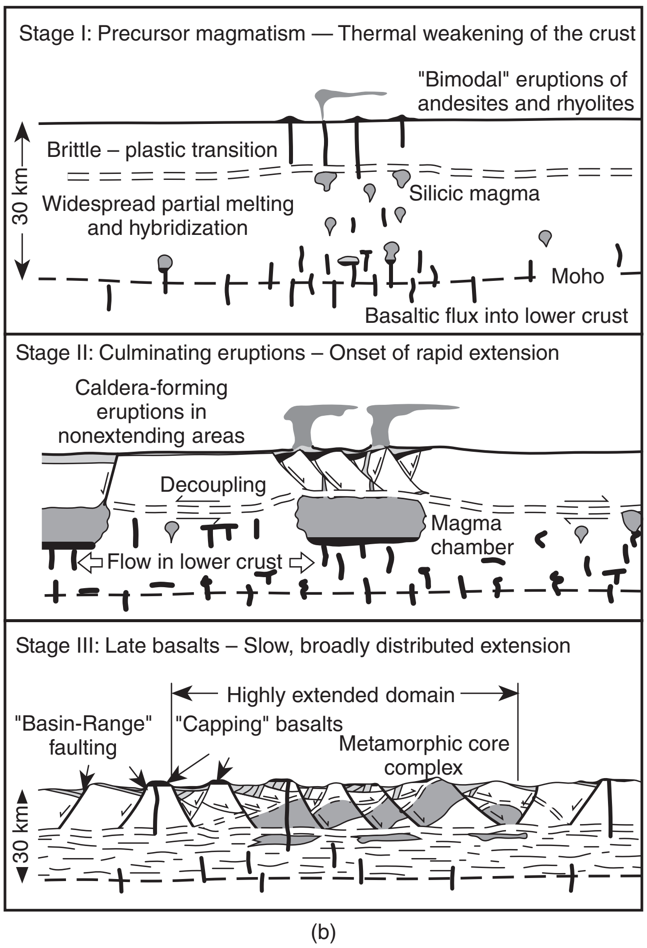


FIGURE 22.1.4 (a) Schematic crustal cross section of the western United States at the end of Mesozoic subduction-related arc magmatism and backarc crustal thickening. (b) Schematic sequence of superimposed Cenozoic Basin and Range extension-related events leading to the present (mostly young) crustal structure of the orogenic belt.

shortening continued uninterrupted into the Early to Middle Tertiary (Figure 22.1.5b). Magmatism in the western U.S. portion of the belt shut off abruptly at about 80 Ma, although subduction continued and, in fact, accelerated, achieving convergence rates of ~15cm/y. These north-to-south differences have been attributed to segmentation of the subducting slab, in which there was an extremely shallow angle of subduction beneath the U.S. portion of the belt. This hypothesis is supported by evidence for rapid cooling of the Sierra Nevada batholith as it moved into a forearc position. As the crust of the arc and backarc was "refrigerated," it regained its rheologic strength and was thus able to transmit stresses for greater distances. During this time, deformation stepped far inboard to Utah, Colorado, and Wyoming, where crustal-penetrating reverse faults caused uplift of the Rocky Mountains during the latest Cretaceous to Eocene Laramide Orogeny (Figure 22.1.5b). Their uplift was contemporaneous with continued shortening in the foreland thrust belt in Arizona and Mexico to the south, and in Montana and British Columbia to the north (Figure 21.1.5b).

Plate motions between the oceanic Kula and Farallon Plates and North America changed again at the end of the Paleocene, and the component of orthogonal convergence diminished rapidly (Figure 22.1.3). In the western United States, it is hypothesized that the shallowly dipping slab either fell away into the mantle or gradually "decomposed" due to conductive heating. Decompression melting of upwelling asthenospheric mantle into the previous region of the slab generated basalts that heated the base of the thickened continental crust, a process that caused extensive assimilation and melting of crustal rocks. This magma mixing ultimately led to eruption of large volumes of intermediate to silicic volcanic rocks (Figure 22.1.5c). Volcanism migrated progressively into the area of previously

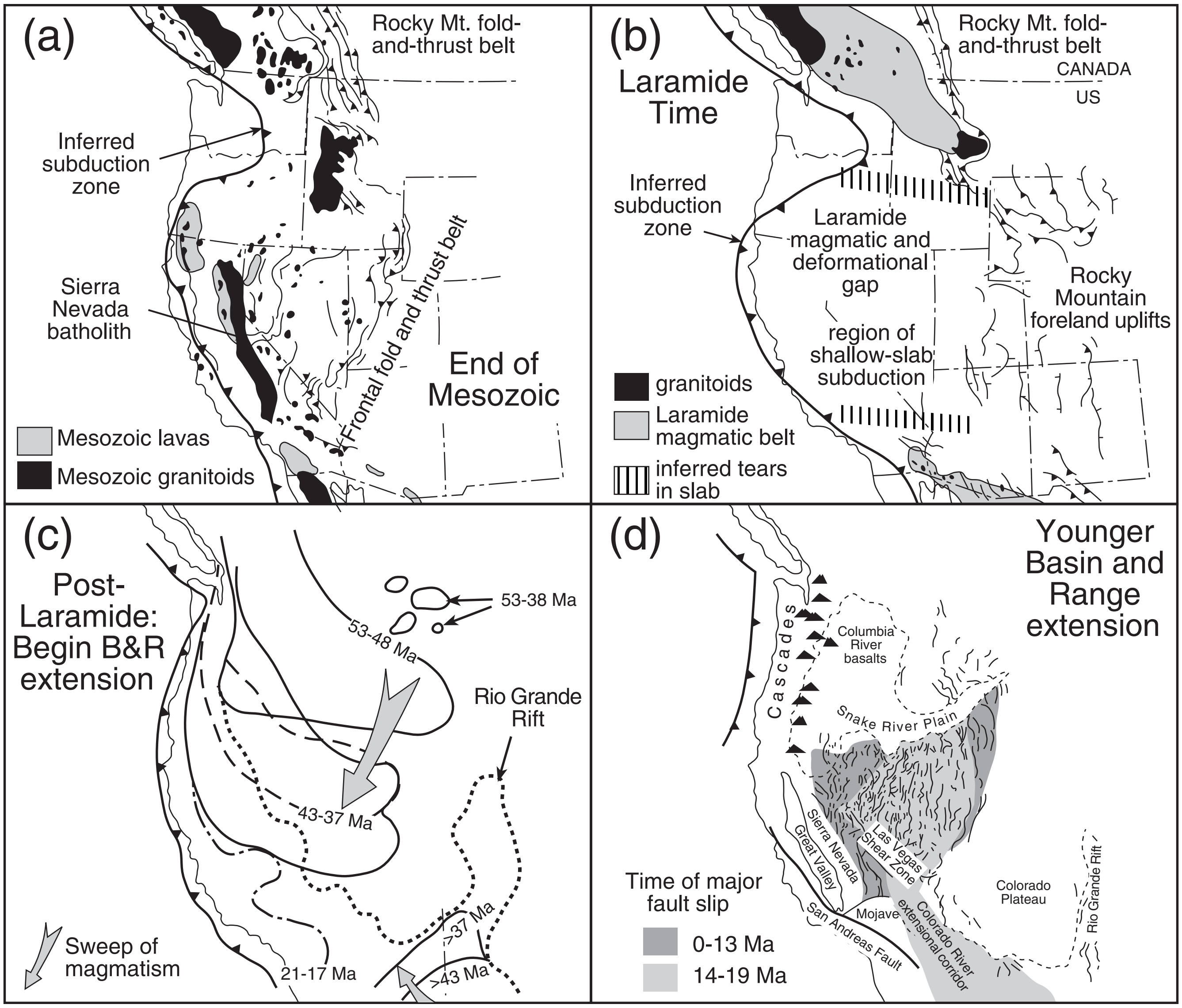


FIGURE 22.1.5 Summary of events leading up to the formation of the Basin and Range Province of the western United States.

flat slab subduction, both southeastwards from the Pacific northwest and northwards from Mexico. The large input of heat into the thick crust rheologically weakened it, and by Miocene time (about 21 Ma), when the slab finally fell away, this heat input triggered wholesale extensional collapse of much of the western United States, resulting in the formation of the present Basin and Range Province (Figure 22.1.5d). This broad zone of continental extension wraps around the southern end of the unextended but (thermally) elevated Colorado Plateau and projects as a finger northwards along the Rio Grande Rift on the eastern side of the plateau. To the west of the Basin and Range Province lies the unextended Sierra Nevada crustal block, with its thicker crustal root, and the virtually undeformed Great Valley sequence, underlain in part by oceanic crust refrigerated during the Mesozoic (Figure 22.1.5d). Volcanism and seismicity are diffuse across this broad zone of continental extension, and thermal springs abound. One of the most impressive physiographic features related to young volcanism is the depression of the Snake River plain, which is believed to represent the Miocene to Recent track of a mantle hot spot that now resides beneath Yellowstone National Park. The present Basin and Range Province, together with associated extension in the Rio Grande Rift and north of the Snake River plain, reflects ~200–300 km of east-west extension that began in the Early to Middle Tertiary and continues today. The modern, regularly spaced basin-and-range physiogeography that lends

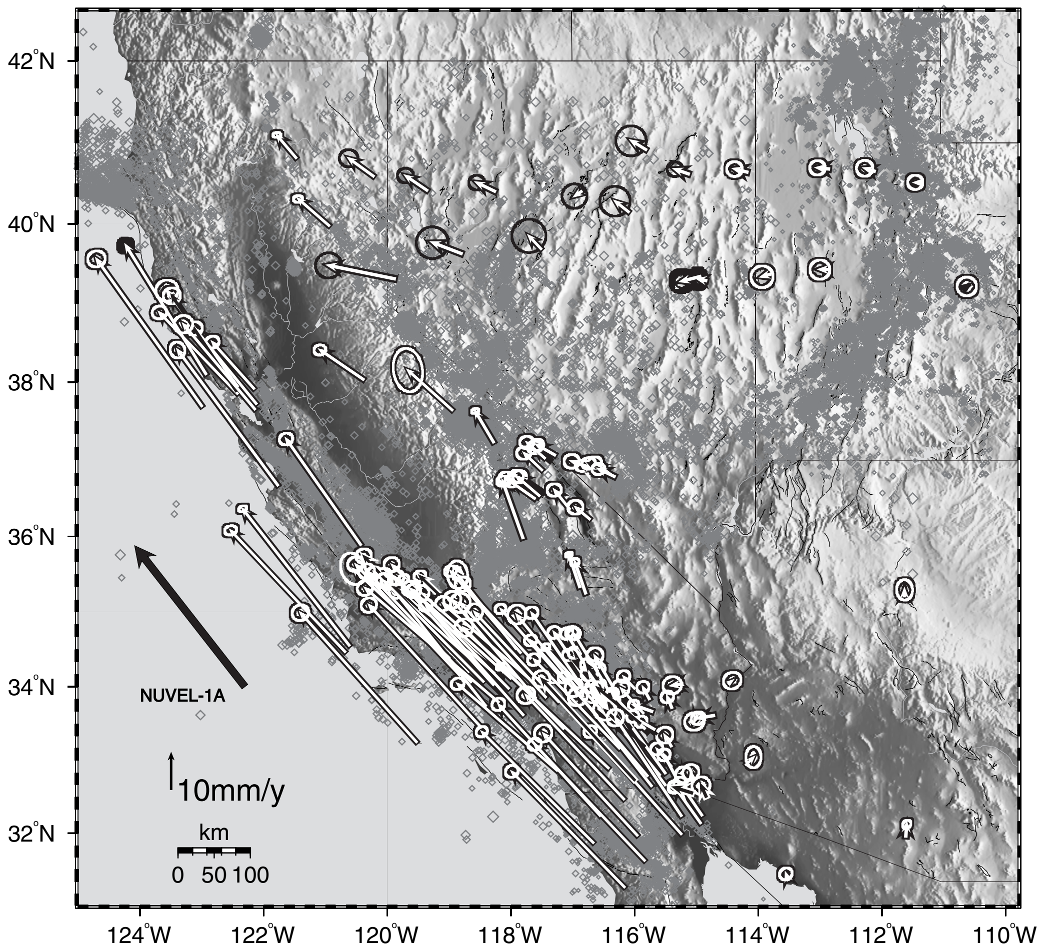


FIGURE 22.1.6 Motion vectors of various sites in the western United States relative to stable North America and NUVEL reference displacement. Site motions show strike-slip along the San Andreas fault system. Extension occurs in the Basin and Range Province north of about 36°E and changes smoothly into the strike-slip motion across a well-defined transition zone. South of 36°E, the San Andreas system accommodates most of the plate motion, and little deformation occurs in the Basin and Range.

the province its name is the surficial manifestation of the youngest system of major normal faults bounding large, tilted, upper crustal blocks (Figure 22.1.5d). Global positioning system (GPS) studies show the partitioning of strain across the western part of the U.S. Cordillera, with some strike-slip motion related to Pacific–North American plate motions taking place in the western part of the Basin and Range Province, between the relatively cold and thick crust of the Sierra Nevada and the hot and thin crust of the Basin and Range (Figure 22.1.6).

Given the long history of ocean-continent plate interaction along the western margin of North America, it may seem surprising that the actual limits and present topography of parts of the Cordillera are dictated mostly by the youngest events to affect the belt (Figure 22.1.1). For example, the Basin and Range Province includes all or parts of the Mesozoic magmatic arc, backarc, and thrust belt, as well as older Paleozoic allochthons and sutures; it is also underlain by the Precambrian rifted western margin of North America. Despite the diversity of tectonic elements across the Basin and Range, the crust is uniformly 25–30 km thick and much of it stands >1 km above sea level, reflecting an anomalously thin and hot mantle lithosphere. The flatness of the Moho across this broad (600 km) extensional province implies that the lower crust was capable of undergoing large-scale flow during extensional deformation (Figure 22.1.4). Thus, it seems clear that the present-day structure of most of the crust and perhaps the entire lithosphere across this region reflects only the youngest event to affect this long-lived orogenic belt. This would imply that if the upper 5–10 km of the crust were removed by erosion, we would probably see very little evidence for the previous 600-m.y. history of this orogenic belt. Convergence presently occurs beneath the Alaskan-Aleutian portion of the margin and beneath the Pacific Northwest, and transform boundaries separate the North American and the Pacific Plates along most of the rest of the margin. In Alaska, shortening and associated diffuse seismicity occur in the overriding North American continent across a broad distance (1000 km). Large-magnitude subduction-zone earthquakes have occurred as recently as 1964 (beneath Anchorage) and uplift by reverse faulting has generated some of the most spectacular and rapidly rising mountains of the Cordillera, including Denali (~6,000 m) in the Alaska Range. Active shortening-related deformation extends northward to the Arctic margin of Alaska. Shortening dies out westward and is replaced by north–south extension in the Bering Strait region, where the mighty Cordillera finally ends. In the Pacific Northwest, folding and thrusting are active in the surficial part of the crust above the Cascadia subduction zone and, as predicted, a subduction-zone earthquake occurred beneath Seattle in early 2001. Detailed studies of contemporary deformation, paleoseismicity studies, and Native oral tradition suggest recurrence intervals of 300 years for such earthquakes, and raise the specter of very large (M>8.0), devastating earthquakes in the future beneath cities such as Portland and Seattle.

In California, the relative motion between the Pacific and North American Plates is partitioned into strike-slip displacement along the San Andreas Fault, and into folding and thrusting related to shortening perpendicular to the San Andreas transform plate boundary (reflected by the recent Loma Prieta and Northridge earthquakes). The exact physical explanation for the observed strain partitioning and how deformation at the surface is coupled with strain at depth in the earth's crust in such zones of continental deformation remain exciting and challenging problems for structural geologists and geophysicists.

22.1.5 Closing Remarks

This brief essay on the geologic and tectonic evolution of the North American Cordillera permit us to make several generalizations about the evolution of such orogenic belts.

- Mountain building (i.e., thickening of continental crust) is not necessarily the result of subduction and collision of allochthonous crustal fragments (terranes) along an active continental margin. Subduction occurred for long spans of time during the history of the Cordilleran margin, but, as in the southwestern Pacific, led mostly to rifting and backarc basin development. True mountain building in the Cordillera appears to have occurred during finite time intervals of rapid convergence and increased absolute westward motion of North America, and was accompanied by magmatic activity.
- The North American Cordillera has long been cited as a classic example of continental growth by the lateral accretion of allochthonous terranes. However, this mechanism is probably not the most fundamental or important process of crustal growth, unless it involves the addition of mature island arcs. Rifting, with formation of rift basins on existing continental shelves, along continental slopes, and within island arcs, and the subsequent filling of these basins by thick prisms of sediment have contributed significantly to the formation of many terranes now incorporated in the Cordillera. Extensional thinning and reworking of continental crust or previously thickened orogenic crust, especially when accompanied by magmatic additions from the mantle, can serve to redistribute and remobilize crust across great portions of an orogen, and the results of these processes often equal or exceed estimates of crustal shortening within the same belts. The best example of this is the reworking and shape-changing of the continent during Cenozoic extension in the western United States.
- Magmatism is a process that is closely linked to deformation in mountain belts. The Cordillera provides excellent examples to illustrate that magmatism is intricately tied to deformation, in that heating causes rheologic weakening of the crust. The rise of magmas transports heat to higher levels of the crust, permitting continents to undergo large-scale deformation, whether by shortening or stretching. This is evidenced by the increasingly better documented eastward migration of magmatism and deformation of the Cordillera in the Mesozoic, as well as by the space-time relation between magmatism and extensional tectonism of the western United States in the Cenozoic.
- Many intriguing questions remain about the evolution of mountain belts such as the Cordillera. Because it is actively deforming, the Cordillera also presents us with a wonderful opportunity to study some of these questions. One of these is how strain-partitioning occurs, that is, where a particular motion vector between two plates or two parts of a continent is partitioned into different styles of deformation in different parts of the orogen (e.g.,

folding and thrusting in the Coast Ranges and strike-slip faulting along the San Andreas Fault in California). Other questions are understanding how seemingly incompatible strains take place within an orogen (e.g., east-west shortening in the Coast Ranges of California and east-west extension in the Basin and Range Province) and what are the driving forces for such strains. The need to answer such questions is a good reason to study contemporary deformations at the scale of the entire orogen. GPS studies measuring contemporary motion across the Cordillera (Figure 22.1.6) are an excellent way to characterize deformations at this scale.

ADDITIONAL READING

The information and ideas in this essay have been distilled from the author's own works and views, and those of many others, as represented in the various chapters of the books, *The Cordilleran Orogen: Conterminous U.S.* and *The Geology of the Cordilleran Orogen in Canada,* two of the volumes of the Geological Society of America's *Decade of North American Geology Series.* A particularly helpful review of the evolution of the western United States is given in Burchfiel et al. (see the following bibliography).

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22.2 THE CASCADIA SUBDUCTION WEDGE: THE ROLE OF ACCRETION, UPLIFT, AND EROSION—An essay by Mark T. Brandon[1]

22.2.1 Introduction

Given the constant surface area of Earth, there must be a balance between the amount of plate created at spreading centers and the amount consumed at subduction zones and other sites of plate convergence. Today, 80% of this return flow occurs at subduction zones, and the other 20% at continent-continent collision zones and diffuse oceanic convergent zones. Modern subduction zones have a total length of 51,000 km, and consume plates at an average rate of 62 km/my (or 62 mm/y), with the highest rates (210 km/my, or 210 mm/y) being found along the Tonga Trench in the southwest Pacific. Subduction zones are marked by Benioff-zone earthquakes and active-arc volcanism, which are indications of shearing and bending of the cold subducting plate and melting caused by dehydration of the plate. Seismologists have generated tomographic images that show subducting plates penetrating deep into the interior of Earth, locally reaching the core-mantle boundary (see Chapter 14).

Subduction zones are not totally efficient in removing the subducting plate. Some fraction of the plate gets left behind as **accretionary complexes** that accumulate at the leading edge of the overriding plate (Figure 22.2.1). In some cases, this accretion might be episodic, involving the collision of large lithospheric blocks, called **tectonostratigraphic terranes.** More commonly, only the sedimentary cover of the downgoing plate is accreted, while the underlying crust and mantle lithosphere are fully subducted. The thickness of this sedimentary cover varies considerably, from hundreds of meters at oceanic subduction zones, like the Mariana system, to as much as 7 km at ocean-continent subduction zones, such as the Makran margin of southwest Pakistan. There is evidence from modern subduction zones that not all of the incoming sedimentary section is accreted. The global rate of sediment subduction has been estimated in one study to be equivalent to an average thickness of 300–500 km of compacted sedimentary rock along all subduction zones combined. This analysis, however, is complicated by deeply subducted sediment that may be accreted at depth beneath the overriding plate, rather than fully subducted into the mantle.

22.2.2 Accretionary Flux

An important theme of this essay is that accretion of largely sedimentary materials has a strong influence on deformation of the overriding plate at subduction zones. We are concerned here with the **accretionary flux** into the subduction zone, which is defined as the thickness of accreted materials times the rate of plate subduction. Sedimentary rocks are quickly compacted during the subduction process, so we use the equivalent thickness of fully compacted sedimentary rock when calculating the accretionary flux. This correction for compaction ranges from ~50% for a thin sedimentary section, which would have a high average porosity, to ~20% for thick sedimentary sections, where the base of the section is already fully compacted.

The Makran subduction zone provides an upper limit for accretionary fluxes at subduction zones. There, the flux could be as high as 210 km^3/my per kilometer of subduction zone (equal to 7 km sedimentary section $\times$ 0.8 compaction factor $\times$ 37 km/my convergence velocity). It is more common to consider the

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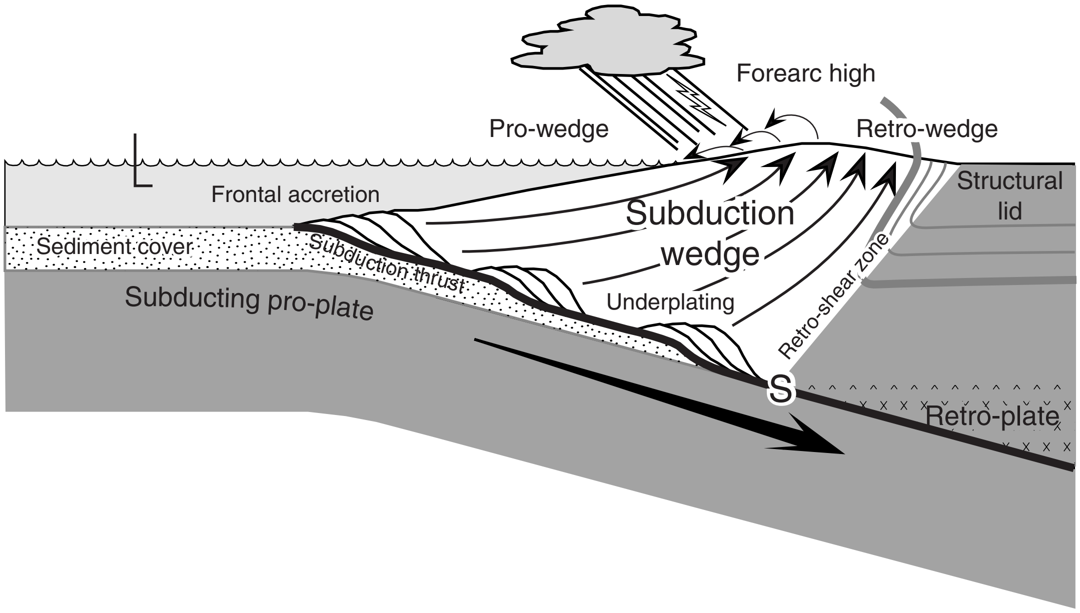

FIGURE 22.2.1 Schematic cross section of a subduction wedge. "S" refers to the S point, the subduction point, where the pro-plate is subducted beneath the retro-plate.

accretionary flux from a cross-sectional view. Thus, the flux is given as km^2/my, which represents the cross-sectional area of material added to the upper plate per unit time.

22.2.3 Wedges, Taper, and Stability

The accreted material tends to accumulate in front of and beneath the leading edge of the overriding plate, forming a wedge-shaped body that grows with time (Figure 22.2.1). A useful analogy is the way that snow piles up in front of a moving plow, forming a tapered wedge that moves with the plow (representing the overriding plate). The wedge entrains snow from the roadbed, which causes the wedge to grow (see Chapter 18).

An important discovery of the 1980s was that accretionary wedges tend to maintain a self-similar form as they grow. This behavior is expected for a wedge made of frictional materials, in that the wedge taper must be blunt enough to overcome the frictional resistance to slip on the subduction thrust that underlies the wedge. A wedge that is able to slip along its base is called a stable wedge, in that the wedge does not deform or change shape as it rides above the subduction thrust. The taper geometry of the wedge introduces an important feedback. Frontal accretion tends to drive the wedge into a substable taper, where the wedge taper is now too slender to overcome friction on the subduction thrust. The wedge becomes locked to the subducting plate, and thus must deform internally to account for the convergence velocity. The resulting horizontal shortening allows the wedge to return to its stable taper, at which point horizontal shortening stops because the wedge is now free to slip on it base. Thus, the accretionary flux into the front of the wedge controls deformation within the wedge.

Erosion will cause similar feedback, as it tends to reduce wedge taper, which causes horizontal shortening and a return to a stable taper. Erosion has a strong influence on the evolution of subaerial convergent wedges, such as the Himalaya or Alps, but subduction wedges are commonly submerged below sea level where erosion is not as active. Nonetheless, erosion can be locally important when a subduction wedge becomes emergent, which is common during the later evolution of wedges at continental subduction zones (e.g., Figure 22.2.1). Erosion acts to drive deformation and also serves to limit the growth of the wedge. This is illustrated by the examples that follow, with particular emphasis on the Cascadia margin of western North America.

22.2.4 Double-Sided Wedges

An important development in wedge theory occurred during the early 1990s, with the recognition that most convergent orogens consist of two wedges, arranged back-to-back. This idea was first proposed for collisional orogens, like the Alps, but it was also recognized as applicable for wedges that form at subduction zones.

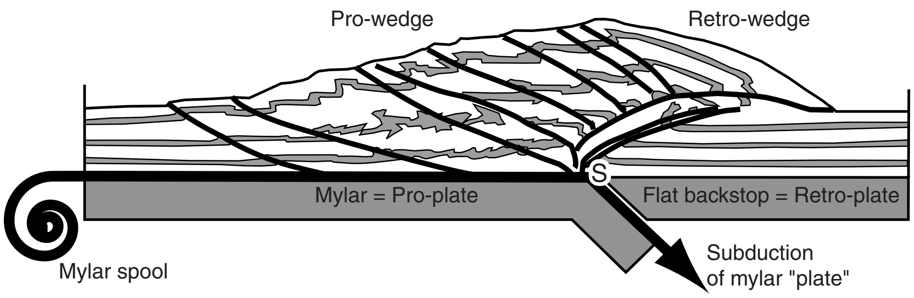

FIGURE 22.2.2 Sketch of sandbox experiment producing a double-sided wedge.

The basic problem with the snowplow analogy was that it was hard to identify where the strong plowblade, or backstop as it is commonly called, was located within the overriding plate. Many authors used relative strength to define backstops within the crust. However, in many cases, the backstop region could be shown to be deforming along with the accreted material in front of it. This contradicted the basic idea of a backstop, which is supposed to be a rigid part of the overriding plate.

The solution to the backstop problem was inspired by sandbox experiments (Figure 22.2.2) in which a mylar sheet serves as the subducting plate and a "fixed" flat-lying plate serves as the overriding plate. This arrangement of "plates" was covered with continuous layers of sand, representing the deformable crust above the plates. A motor was then used to draw the mylar downward through a slot in the base of the sandbox, thus simulating the motion of the subducting plate. This subduction caused the overlying sand layer to deform into a double-sided wedge centered over the subduction point, or S point. The overriding plate served as a backstop, but in this case the backstop was flat-lying. Notably, there is no visible backstop at the surface. Instead, the backstop is a deep-seated structural element, with a flat-lying geometry that is hidden from view.

In a geodynamic perspective, deformation within a convergent wedge is driven by the motion of a subducting pro-plate and an overriding retro-plate. These two plates are rigid, and correspond to relatively strong lithospheric mantle, whereas the overlying crust is deformable, and thus must accommodate the velocity discontinuity at the S point. In simple terms, the model postulates that convergent wedges overlie a deep-seated mantle subduction zone. It is the cold strong mantle in the plates that controls their plate-like behavior. The crust is draped over the zone of mantle subduction, and thus deforms in a more distributed matter to accommodate subduction.

Wedges associated with subduction zones are called **accretionary wedges,** but the term has led to the thinking that the size of an actively deforming wedge is defined by the volume of accreted sediment. The concept of a double-sided wedge provides a different view, in that the active wedge involves both accreted sediments and upper-plate rocks. This result stems from the fact that the width of the wedge is determined by the motion of the underlying mantle plates, and not by any specific distribution of strength in the crust. Using terms like **subduction wedge** or **convergent wedge** when describing double-sided wedges helps to distinguish this model from that of a small wedge of accreted sediment bounded by a strong crustal backstop.

22.2.5 Subduction Polarity and Pro-Side Accretion

The term facing is used to describe the polarity of subduction. For instance, a west-facing subduction zone indicates that the overriding plate is moving westward relative to the subducting plate. This asymmetry can be ignored when measuring convergence velocities across a subduction zone, but it does have an important influence on how the wedge grows with time. Consider the situation where the pro-plate is subducted, while the retro-plate remains largely unconsumed. This asymmetry means that most of the accretionary flux is carried into the pro-side of the convergent wedge. When accretion occurs on the retro-side as well, it is always at a much slower rate than on the pro-side.

The dominance of pro-side accretion creates a strong tendency for all material in the wedge to move horizontally from the pro-side of the wedge to the retro-side. This situation is further influenced by the pattern of surficial erosion across the wedge. Consider accretion on the pro-side and strong erosion on the retro-side. This pattern of accretion and erosion will produce the greatest horizontal velocities within the wedge. An example of this situation is the Southern Alps of New Zealand, a beautiful active mountain range formed by oblique subduction of the Pacific Plate beneath the Australian Plate. The retro-wedge

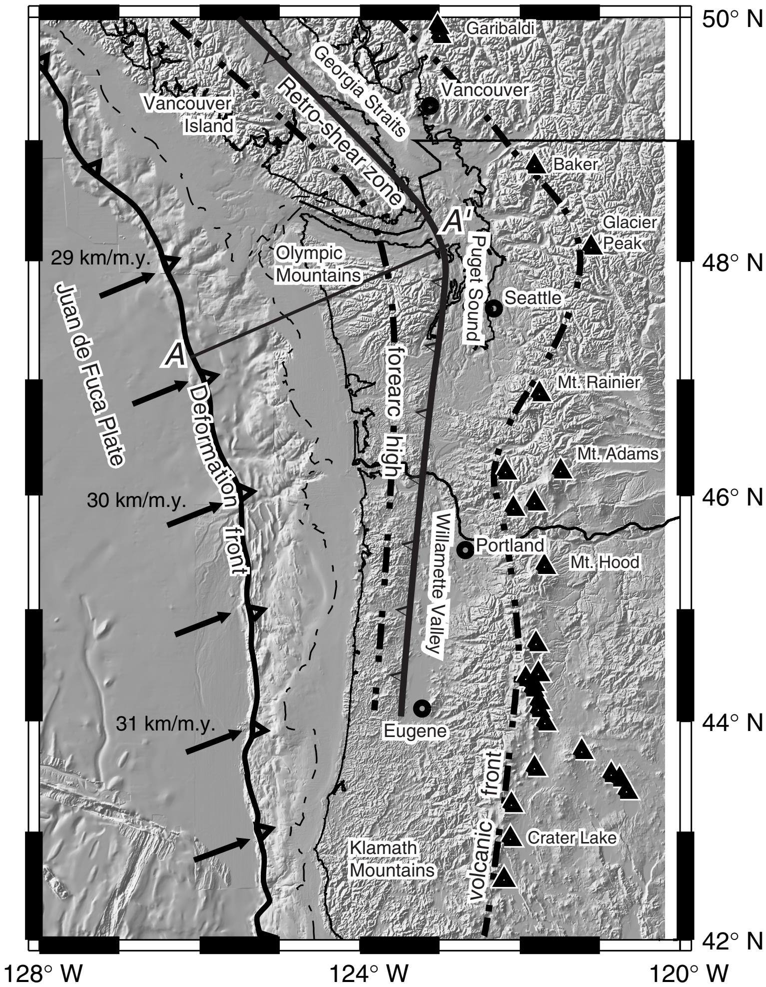

FIGURE 22.2.3 Modern tectonic setting of the Cascadia margin, emphasizing the plate boundary, volcanic arc, and physiography of the modern margin. *A-A′* shows the location of the cross section in Figure 22.2.4. Velocity vectors show the modern velocity of the Juan de Fuca Plate relative to a fixed overriding plate. Recent geodetic work has shown that the Cascade arc and forearc move as a separate plate relative to North America, due in part to deformation across the Basin and Range Province, which lies behind the southern half of the Cascade arc. The velocities shown here at the subduction zone account for this additional plate.

lies on the west side of the South Island, and is subjected to big storms coming from the Tasman Sea. Average precipitation there is more than 7 m per year, whereas the pro-wedge, located on the drier east side of the range, has an average precipitation of ~1 m per year. As a result, erosion associated with those Tasman Sea storms tends to move material from the pro-side to the retro-side of the wedge, over a horizontal distance of ~90 km.

At this point it is useful to distinguish two important geologic features within the wedge system: the structural lid and the accretionary complex. The structural lid refers to the tapered leading edge of the overriding plate, which tends to get involved in wedge deformation. The accretionary complex refers to those materials that were accreted into the wedge. The important distinction is that the structural lid originated as part of the retro-plate, whereas most, if not all, of the accretionary complex is derived by accretion from the pro-plate. From the examples that follow, we will see that, as the accretionary complex grows, the structural lid tends to get shouldered aside to the rear of the wedge. A common result is that the structural lid is only preserved within a large backfold within the retro-wedge.

22.2.6 The Cascadia Subduction Zone

The Cascadia subduction zone has a length of 1300 km and is located along the continental margin west of northern California, Oregon, Washington, and Vancouver Island (Figures 22.2.3 and 22.2.4a). In this area, the offshore Juan de Fuca Plate is actively subducting beneath an overriding North American Plate at rates of ~30 km/m.y. (arrows in Figure 22.2.3; see caption for details on estimation of convergence velocities). The direction of convergence is approximately orthogonal to the subduction zone. The surface trace of the subduction thrust is entirely submarine (deformation front in Figures 22.2.3 and 22.2.4a), lying at ~2500 m below sea level. Most of the plate convergence is accommodated by slip on the subduction thrust, but there is also a subordinate amount of shortening in the overriding subduction wedge. Slip on the subduction thrust is thought to occur episodically during great thrust

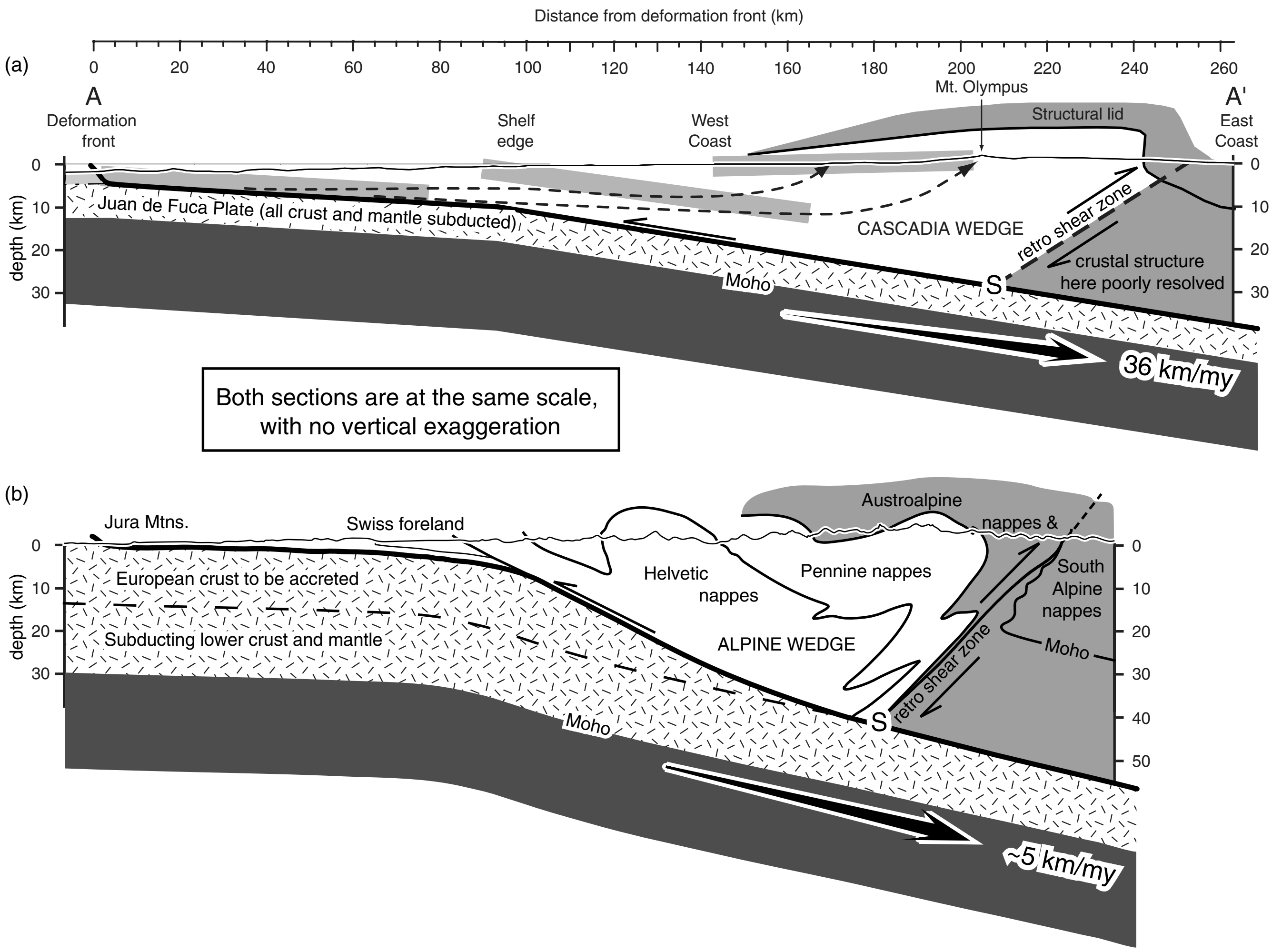

FIGURE 22.2.4 Comparison in cross section of the structure of the Cascadia margin across northwest Washington State and the structure of the European Alps across central Switzerland. The gray bands in the Cascadia section illustrate the displacement history of sediments presently exposed in the western Olympics, which were accreted at the front of the wedge at ~22 Ma and then moved rearward through the wedge, reaching the west side of the Olympics in the present.

earthquakes, with recurrence times of ~500 years. The last subduction-zone earthquake occurred on AD January 26, 1700. The Juan de Fuca Plate is covered by ~2500 m of sediment. Geophysical and geochemical evidence indicates that all of the incoming sedimentary section is accreted into the Cascadia wedge, whereas the underlying crust and mantle are fully subducted. These observations allow an estimate of the present accretionary flux of ~52 km^2/my.

The subduction zone parallels the Cascade volcanic arc, which includes active volcanoes such as Mt. Rainer. Studies at modern subduction zones indicate that the subducting slab has to reach a depth of ~100 km for melting to start in the overlying mantle. Thus, the Cascade volcanic front can be viewed as an approximate indicator of the 100-km depth contour of the downgoing slab. With this in mind, note that the distance between the subduction zone and the arc is largest across the Olympic sector of the Cascadia margin. The reason is that the dip of the subducting slab varies along the subduction zone and that the shallowest dip is found beneath the Olympic Mountains. We return to this point subsequently because the shallower dip of the slab beneath the Olympics has had a strong influence on the evolution of the Cascadia wedge in that area.

The forearc, which is the region between the arc and the subduction zone, is marked by a low and a high, both of which parallel the general trend of the subduction zone. The forearc low includes the Georgia Straits adjacent to Vancouver Island, Puget Sound of western Washington State, and the Willamette Valley of western Oregon. The forearc high corresponds to the Insular

Range, which defines Vancouver Island, and the Coast Range of western Washington and western Oregon.

The forearc high represents the crest of the Cascadia subduction wedge, with a pro-wedge to the west and a smaller retro-wedge to the east. The retro-wedge shear zone, shown in Figures 22.2.3 and 22.2.4a, marks the eastern limit of permanent deformation and uplift associated with the subduction wedge. The structure exposed there is not a fault zone, but rather a large eastward vergent "backfold," which is actively accommodating slow, top-to-east shear between the retro-wedge and the forearc low. The forearc low is commonly viewed as a basin, but there has been little subsidence or sediment accumulated in this low over the last 10 to 15 my. Instead, the low is defined by the volcanic arc to the east and the actively growing forearc high to the west.

Geologic evidence indicates that the Olympic Mountains mark the first part of the Cascadia forearc high to emerge above sea level, at ~15 Ma. The forearc high apparently grew more slowly elsewhere along the margin. In fact, Vancouver Island and the southern part of the Coast Ranges may have emerged above sea level only within the last 5 to 10 my. The early emergence of the forearc high in the Olympics may reflect a slab that is shallower there. To understand this situation, consider that the growth of a subduction wedge is controlled by the accretionary flux into the wedge and erosion of the forearc high. Erosion cannot start until the forearc high becomes subaerially exposed, but once the wedge does become emergent, erosion will act to slow the growth of the wedge, until it reaches a flux steady state, when the erosional outflux becomes equal to the accretionary influx.

It was already noted that the subducting slab has a shallower dip beneath the Olympics relative to adjacent parts of the subduction zone. This archlike form of the subducting slab is apparent when one considers the depth of the slab beneath the forearc high, which is equal to ~40 km beneath Vancouver Island, ~30 km beneath the Olympics, and ~40 km beneath southwest Washington State. This variation in slab depth cannot be attributed to variations in topography, because the mean elevation of the forearc high varies by only a small amount, between ~500 to 1000 m along its length. Thus, this configuration may be a fundamental feature of the subduction geometry of the slab.

The important point here is that the wedge beneath the Olympics is much smaller than elsewhere along the margin. Thus, given a similar accretionary flux, less time was needed for the forearc high to become emergent in the Olympics relative to other parts of the subduction wedge.

Erosion of the Forearc High

The effect that erosion has on deformation and growth of the wedge was noted earlier. Several methods were used to measure the long-term erosional flux coming out of the Olympics part of the forearc high. One method is **apatite fission-track dating,** which dates when a sample cooled below the apatite fission-track closure temperature (~110°C). A thermal model is used to convert the closure temperature to a depth, with results that typically lie in the range of 3–5 km. Dividing depth by cooling age gives the average erosion rate for the time interval represented by the cooling age.

Another method involves measuring the **incision rate** of rivers, which is the rate at which a river cuts downward into bedrock. The rate of channel incision can be determined from old river deposits preserved on the hillslopes adjacent to the river. These deposits commonly overlie old remnants of the river channel where it was running on bedrock. The height of these paleochannel remnants above the modern channel divided by the age of the paleochannel gives the incision rate of the river.

In the Olympics, paleochannel remnants formed at ~65 ka and 140 ka produced 60 incision rates along the Clearwater River, which drains the west side of the forearc high. Apatite fission-track ages provide another 43 estimates of erosion rates, with average cooling ages of ~7 my. These data are shown in Figure 22.2.5, where they have been projected into the section line *A-A′* (Figure 22.2.4) across the subaerially exposed part of the forearc high. Particularly interesting is the similarity between incision rates and fission-track-based erosion rates, despite the fact that they represent local versus regional processes and different intervals of time as well; that is, 100 ky versus 7 my. Recently (U-Th)/He apatite dating found younger cooling ages of ~2 my, which reflect the lower closure temperature (~65°C) for this thermochronometer. The erosion rates indicated by these ages (not shown) also match closely the rates shown in Figure 22.2.5. The conclusion is that, when viewed on timescales longer than ~50 ky, the pattern and rates of erosion across the forearc high appear to have been fairly steady.

Let's compare these rates with the accretionary flux that we estimated previously. We use the data in Fig-ure 22.2.5 to estimate the integrated erosional flux from the forearc high. The curve provides a smoothed version of the data and is used to integrate the flux. This gives an erosional flux of 51 km^2/my. Our thermal-kinematic modeling gives a more precise estimate of 57 km^2/my. Both estimates are similar to the accretionary flux, 52 km^2/my, estimated

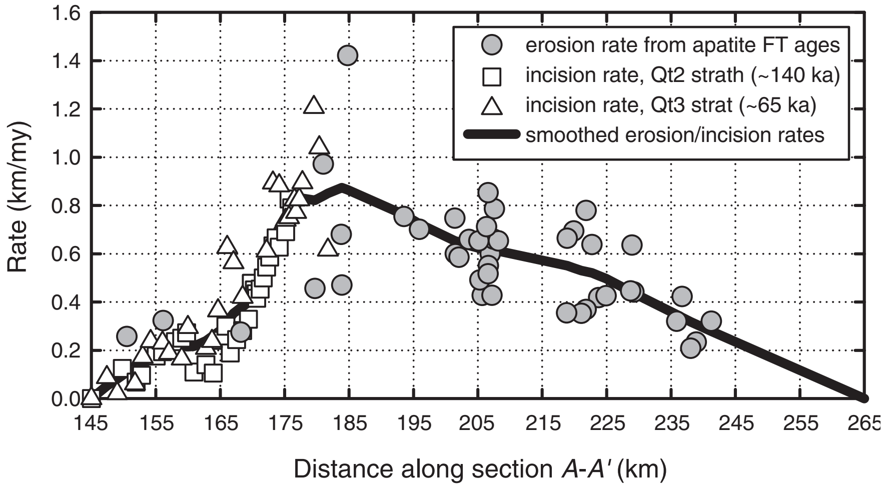

FIGURE 22.2.5 Fluvial incision rates and long-term erosion rates determined from apatite fission-track ages. The fluvial incision rates represent downcutting of the Clearwater River over a timeframe of ~100 ka, whereas the fission-track ages indicate erosion rates over a time frame of ~7 my. The similarity in rates suggests that the pattern and rates of erosion have been steady across this part of the forearc high. The black curve represents a smoothed version of the erosion rate distribution across the high. Integration of this curve indicates that the long-term erosional flux from the forearc high is ~51 km^2/my.

previously, showing that there is a close balance between accretionary flux into the wedge and the erosional flux out of the wedge. Thus, the wedge cannot get any larger with time. Even so, the material within the wedge will still continue to move and deform, given that an eroding wedge must thicken to maintain a stable taper.

(U-Th)/He apatite dating in other parts of the Cascadia forearc arc high shows that everywhere else the forearc high has been only slightly eroded, by less than 3 km. Estimates of erosion rates suggest that these regions have only recently become emergent and have yet to reach the steady-state configuration observed in the Olympic Mountains. It would be interesting to predict how much time is needed for a growing subduction wedge to reach a steady-state configuration. To do so requires a better understanding of how local-scale erosional processes scale up in time and space to account for the large-scale evolution of orogenic topography. This is an area of active research, with much promise for new discoveries.

DEFORMATION AND EROSION OF THE STRUCTURAL LID

How have accretion and erosion influenced the longer-term evolution of the Cascadia wedge? Prior to initiation at ~35 Ma, the continental margin included a large oceanic terrane, Eocene in age, called the Crescent terrane. This terrane was already part of the outboard edge of the North American Plate, having been accreted to North America prior to 35 Ma. Thus, when the Cascadia subduction zone was initiated, the Crescent terrane became the structural lid for the subduction zone (Figure 22.2.4a).

Uplift and erosion in the Olympics provide a good sectional view of the Crescent terrane (Figures 22.2.4a and 22.2.6). It consists of relatively coherent Eocene oceanic crust, mostly pillowed flows of basalt, which are typical of submarine volcanism. Exposures along the north side of the Olympic Mountains show the tapered form of the structural lid (Figure 22.2.6). Across the 120-km width of the forearc high, the basaltic sequence changes from 4 km thickness in the west to more than 15 km in the east. This tapered geometry marks the original eastward dip of the Cascadia subduction zone when it was first formed. At that time, the lower part of the Crescent terrane was apparently subducted, leaving behind the tapered lid, which then became the highest structural unit within the Cascadia wedge.

In the Olympic Mountains, the structural lid is underlain by an accretionary complex, called the Olympic subduction complex. In cross section, this unit is 240 km across and extends to a depth of at least 30 km, giving it a cross-sectional area of 3600 km^2. Seismic studies indicate that this accretionary complex is made up entirely of sedimentary rocks; there is no evidence of basaltic crust or mantle from the subducting plate. Seismic reflection profiles across southern Vancouver Island indicate that the Crescent terrane there is also underlain by a large volume of low-velocity layered rocks, similar to the Olympic subduction complex. Given the present accretionary flux, it would take ~70 my to grow an accretionary complex this size, which is much longer than the 35 my age of the Cascadia subduction zone. Evidence in the Olympics indicates that the more internal parts of the accretionary complex are made up of sedimentary sequences that probably predated the subduction zone. This makes sense, given that the newly formed subduction zone cut across a preexisting continental margin. Thus the sedimentary units that flanked that margin were probably the first to be accreted.

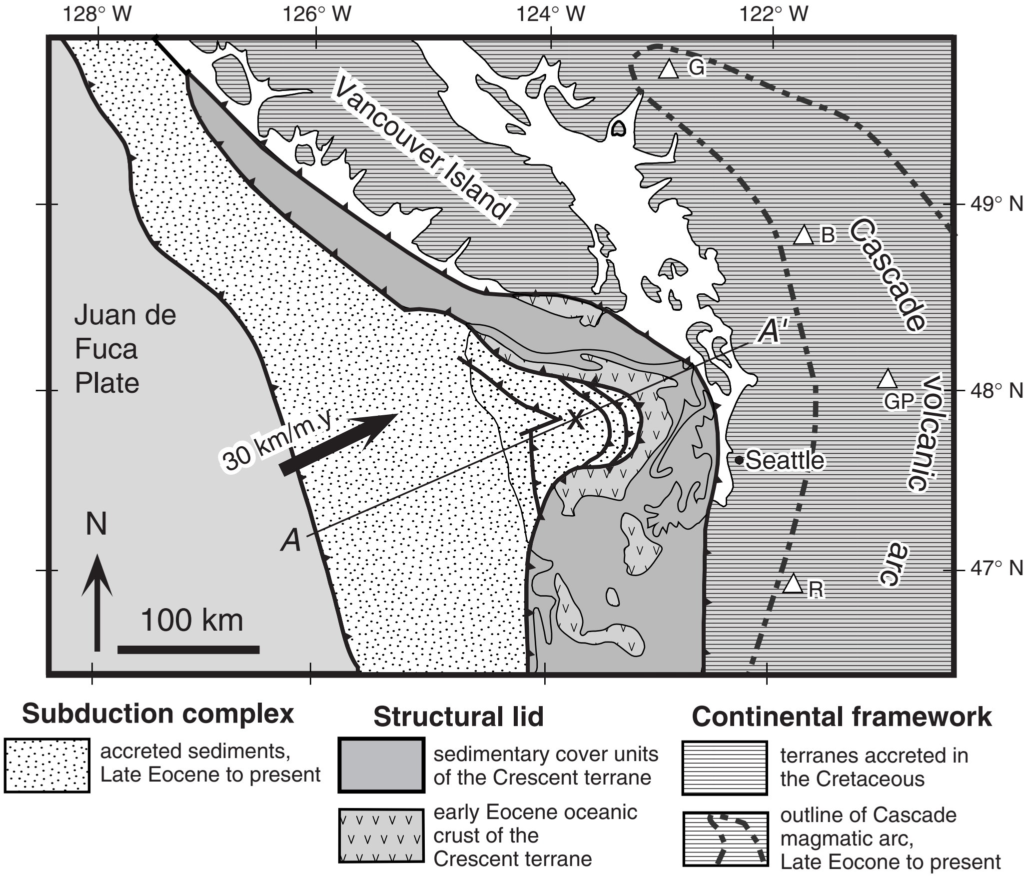

FIGURE 22.2.6 Geology of the Cascadia margin, emphasizing the present configuration of the structural lid. "X" marks Mount Olympus, which is the highest summit (2,428 m) in the Washington-Oregon Coast Range. *A-A′* indicates the location of the cross section shown in Figure 22.2.3. Major volcanoes of the Cascadia arc are: G = Mt. Garibaldi, B = Mt. Rainer, GP = Glacier Peak, R = Mt. Rainier.

In the Olympics, the structural lid has slowly been driven into the back of the wedge system as the accretionary part of the wedge has grown in size. Initially, the lid was uplifted into a broad arch, as observed in the Oregon Coast Range. When this arch became emergent, it started to erode, which allowed the lid to rise faster. As the lid was removed, the accretionary complex moved rearward to take its place. This motion was required for the wedge to maintain a stable taper, as discussed earlier.

The end result is that the lid was driven rearward, creating a large fold. This fold is actually the manifestation of a broad, west-dipping retro-shear zone, which accommodates the top-east motion of material within the retro-wedge. This structure is quite impressive in the Olympics. One can drive into the mountain range from the east, where the Crescent terrane is still flat-lying and undeformed. Moving west across the range front, one notices that the Crescent starts to dip to the east, which marks the fact that we have crossed into the upper limb of the back fold. The dip continues to increase until the section becomes vertical, and locally overturned. This increase in limb dip is also accompanied by a dramatic topographic gradient, which marks the east-dipping surface of the retro-wedge. Within 20 km, the topography rises from sea level to over 2300 m.

The Olympics represents the most evolved part of the Cascadia forearc high, as indicated by uplift, erosion, and backfolding of the structural lid. Elsewhere along the Cascadia margin, the structural lid still covers the forearc high. This difference in evolution of the margin accounts for the large reentrant in the Crescent terrane centered on the Olympic Mountains. Other parts of the forearc high continue to grow, but that growth has only resulted so far in a broad arching of the structural lid. The prediction is that with further deformation and erosion, those other parts of the forearc high will evolve towards the structural configuration observed in the Olympics.

22.2.7 Comparison between the Cascadia and Alpine Wedges

The cross sections in Figure 22.2.4 offer a comparison between the Cascadia wedge and Europe's Alpine wedge (see Subchapter 21.1). The Swiss Alps are different in that they formed by collision of two continental plates, starting ~50 Ma. Convergence there has slowed down, and perhaps stopped within the last 5 my The average Alpine convergence rate is ~5 km/my, which is slower than the Cascadia convergence by a factor of six. However, in the European Alps, the pro-thrust cuts much more deeply into the pro-plate, resulting in accretion of the upper 15 km of crust from the pro-plate (Figure 22.2.4b). The associated accretionary flux for the Alps is therefore estimated ~75 km^2/my, which is slightly larger than for the Cascadia wedge. Nevertheless, the Alpine wedge is smaller than the Cascadia wedge, but because the Alpine wedge is entirely subaerial, erosion was able to more effectively limit the size of the wedge.

This comparison highlights the similarities in backfolding of the structural lid. In the European Alps, the structural lid is made up of crustal rocks belonging to the Southern Alpine and Austroalpine nappes (Section 21.1). These nappes are thrust sheets associated with the tapered leading edge of the overriding continental plate, which is called the Adriatic Plate. As in the Olympics, the tapered geometry of the structural lid reflects the cut made through the upper plate by the subduction thrust when subduction was initiated. Accretion from the pro-side is responsible for driving the structural lid rearward in the wedge. The Alpine geologists call this retrocharriage, meaning "to carry back." The resulting backfold underlies the southern flank of the Alps.

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22.3 THE CENTRAL ANDES: A NATURAL LABORATORY FOR NONCOLLISIONAL MOUNTAIN BUILDING—An essay by Richard W. Allmendinger and Teresa E. Jordan[1]

22.3.1 Introduction

One of the great early advances of plate tectonics was the realization that mountain building is associated with activity at the margins of tectonic plates. The **Andes** represent the case of mountain building produced by the convergence of an oceanic and a continental plate, a relatively simple and elegant end member in the spectrum of orogenesis (Figure 22.3.1). The other extreme is represented by continent-continent collision, for example the Himalayan-Tibetan system. Because of the association of the Andes with andesites and the chain of volcanoes that ring the Pacific, a widespread misconception among geologists outside of South America has been that the Andes Mountains are primarily a volcanic edifice. In fact, it is now clear that the crustal thickening that produced the Andes is mostly structural in origin and that the volcanoes for which the mountain belt is best known sit on top of that structural welt.

During the 1980s, many investigators sought evidence of accretionary events in the Andes that might have been responsible for Andean mountain building. Underlying this search was the concept that a major mountain system produced by horizontal contraction must be produced by collision of two buoyant crustal masses. However, nearly a decade of paleomagnetic study in Chile and Peru has turned up no evidence that accretion played even a minor role in building the modern Andes. Instead, the evidence suggests that material was removed from the continental margin during the Andean orogeny.

Thus, in the central Andes mountain building occurs by dominantly structural processes in a noncollisional setting, in which the oceanic Nazca Plate is subducted beneath the continental South American plate (Figure 22.3.2). The purpose of this essay is to describe these processes, as well as the general tectonic setting of the Andes. Several factors make the central Andes, located between 5° and 35°S latitude, a unique laboratory of orogenesis. Because the deformation is active today, the governing boundary conditions can be identified and, in some cases, quantified. These include (1) the plate convergence rate and obliquity, (2) the geometry of the subduction zone, and (3) the dynamic topography, which reflects the interaction of tectonic and climatic processes. Furthermore, crustal seismicity gives us an idea of short-term strain rates, as well as the distribution of modes of failure in the continental crust.

The modern Andes are commonly regarded as a recent analog for ancient mountain belts, such as in the Mesozoic-Early Tertiary Cordillera of the western United States (Figure 22.3.3). Although the modern setting is quite simple, western South America has a complex pre-Andean history. Thus, the starting materials for Andean deformation were extremely heterogeneous and their responses to the stresses that produced Andean deformation are equally varied. This factor becomes particularly important when one tries to decide whether a particular structural geometry owes its existence to the modern plate setting, or to ancient anisotropies of the continental crust.

22.3.2 The Andean Orogeny

Subduction of oceanic crust beneath western South America has occurred more-or-less continuously since the Middle Jurassic. The term "Andean Orogeny" refers collectively to all tectonism that occurred since the Jurassic. Although subduction has been continuous, the

[1]Both at Cornell University, Ithaca, NY.

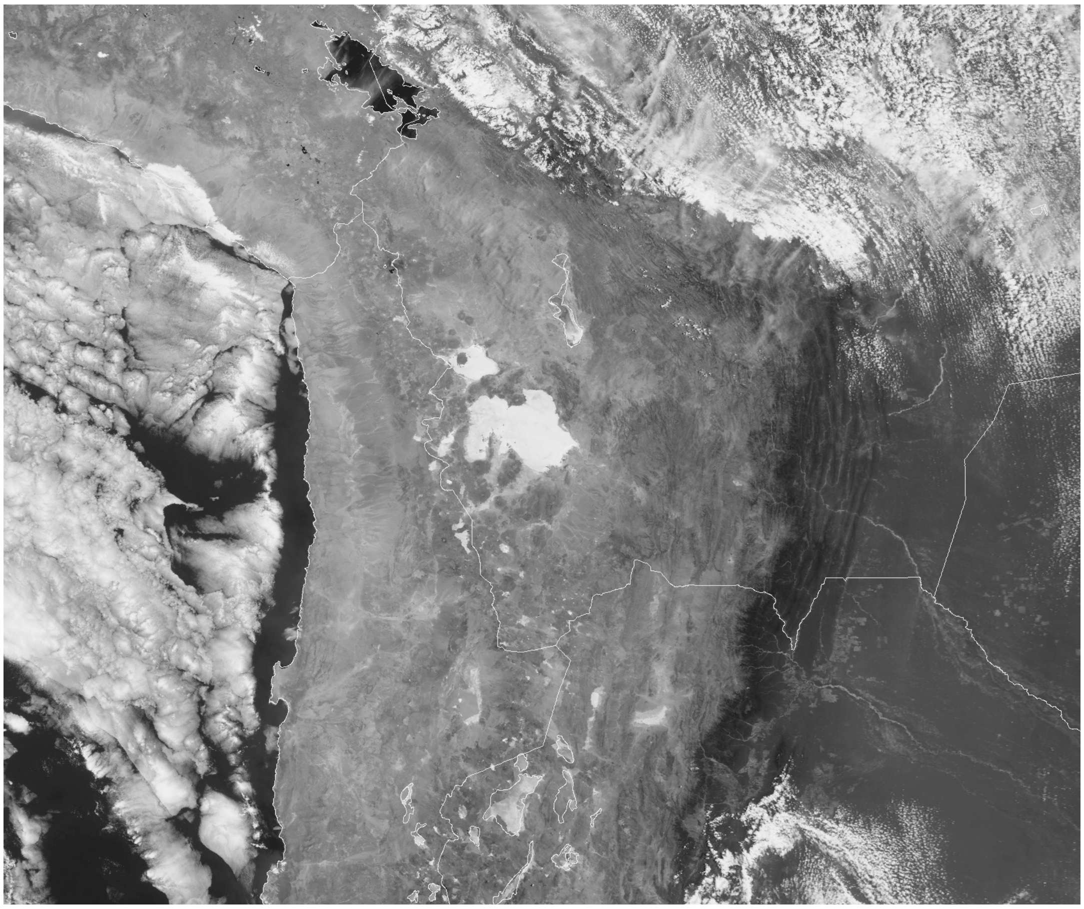

FIGURE 22.3.1 Landsat image of the central Andes in South America.

obliquity and rate of convergence, as well as the dip of the subducted plate, have varied considerably. Thus, the style and distribution of mountain building have not been continuous or uniform. Here we focus mostly on structures developed during the last 30 m.y. of the Andean Orogeny, but we begin by briefly reviewing the older events.

Arc-related igneous rocks of Jurassic and Cretaceous age occur along the present coastline in Chile and Peru and, as one moves progressively farther inland (i.e., eastward), the arc becomes progressively younger. The Mesozoic magmatic arc is anomalously close to the present-day trench (within 75 km in some cases), leading to the conclusion that the Mesozoic forearc region has been stripped off the margin and has either been subducted or perhaps moved laterally.

In many areas, the first significant deformation of continental crust associated with Andean subduction was horizontal extension. Based primarily on interpretation of lithologic associations, the Jurassic and Cretaceous has long been suspected as a time of intra-arc or backarc rifting. Both in the northern Andes of Colombia and Ecuador and in the southernmost Andes of Tierra del Fuego, this rifting culminated in the production of new oceanic crust. In the central Andes, the clues are more subtle, but work in northern Chile at 27°S has yielded unique data on the geometry and kinematics of the extensional structures of this event. Those structures, low-angle normal faults, extensional chaos, domino blocks, and so on, are geometrically comparable to those seen in Cenozoic detachment terranes of the western United States.

Early Tertiary Andean orogenesis is commonly referred to as the "Incaic Orogeny." This phase of deformation is largely restricted to the present forearc in Chile and Peru. However, deep erosion on the eastern side of the central Andes northeast of La Paz, Bolivia, has revealed rocks that were metamorphosed (or cooled through about 300°C) at about 40 Ma, the same time as mountain building occurred farther west.

The morphologic edifice that we associate with today's Andes is a product of mountain building during just the last 30 m.y. Most of the surface-breaking structures associated with this phase of deformation are concentrated within the high topography and on the eastern side of the mountain range. Many workers refer to this young phase of deformation as the "Quechua Orogeny," although different workers use

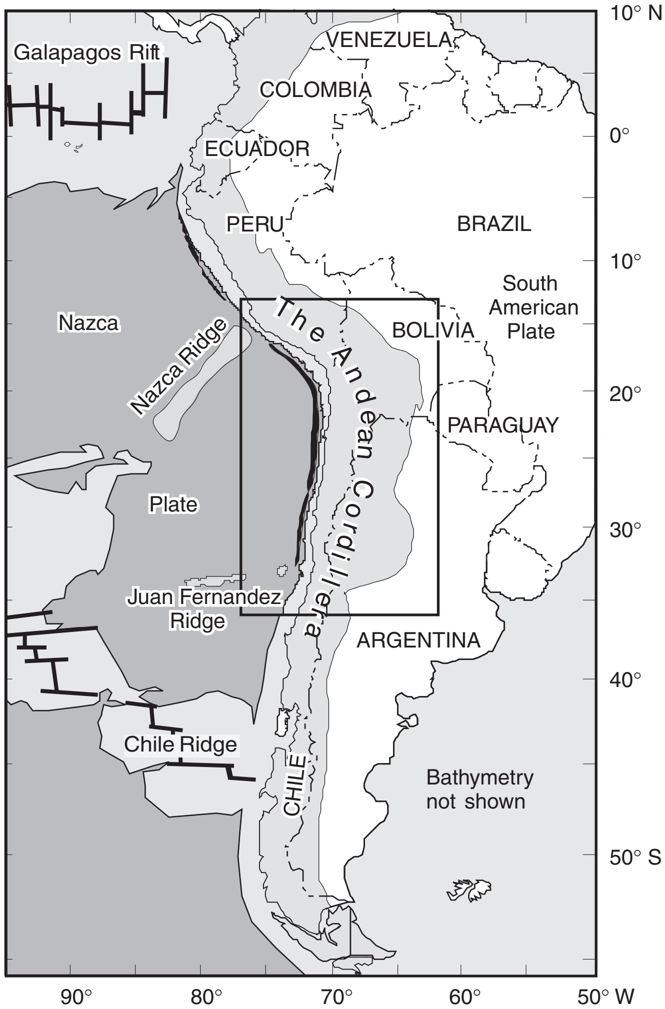

FIGURE 22.3.2 Generalized map of the Nazca-South American plate boundary showing the principal bathymetric features of the Nazca Plate and the extent of the Andean Cordillera. Bathymetric contours shown at –3500 m and –5500 m; the latter, with depths greater than –5500 m shown in black, marks the position of the Peru-Chile Trench. The box shows the position of Figure 22.3.3.

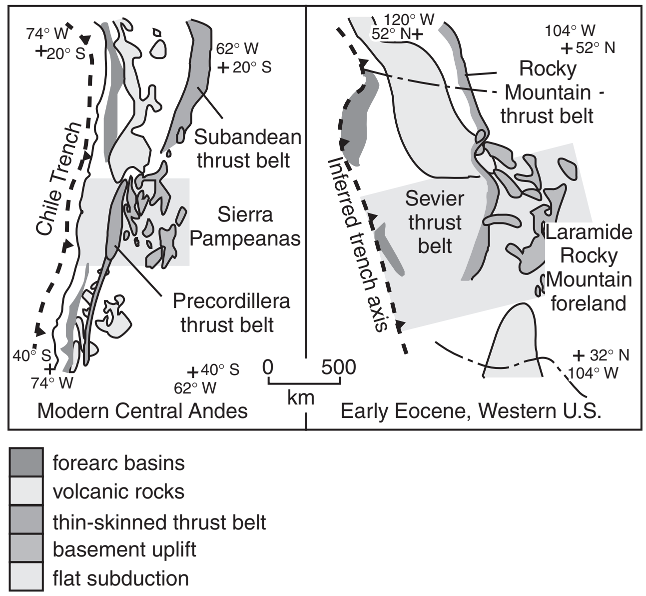

FIGURE 22.3.3 Comparison of the major tectonic elements of the modern central Andes (left) and the early Eocene of the western United States (right) both at the same scale.

this term to refer to events of different ages. The topography of the Andes is the result of structural and thermal processes and is not simply due to piling up of volcanic rocks; in the central Andes, deformed Paleozoic to Cenozoic rocks are commonly found as high as 4 km or more, where they form the great bulk of the high plateau known as the Altiplano. Miocene to recent volcanoes are perched on top of this plateau.

22.3.3 Late Cenozoic Tectonics of the Andes

The Nazca Plate is currently being subducted beneath western South America in a direction of 077±12° at about 10 cm/y, a rate that varies little along the strike of the plate boundary from the triple junction at 49° S to at least 0° of latitude in Ecuador (Figure 22.3.4). Nonetheless, the geometry of the subducting slab is highly variable for reasons that are not totally understood. Between the triple junction and ~33° S, the slab dips ~30°E, but can only be tracked to a position directly beneath the modern arc; subduction zone earthquakes beneath the backarc region are virtually unknown. From 33° to 28°S, the subducted plate dips at ~30° down to depths of ~100 km, but it is nearly flat farther east; nearly 600 km east of the plate boundary beneath the city of Córdoba, Argentina, the subducted plate is only about 200 km deep. This segment of the Andes has had no volcanism since the late Miocene. To the north of 28°S, the plate gradually steepens and, by ~24 S, has resumed its uniform 30°E dip. This central steep-dipping segment correlates with the Central Volcanic Zone of the Andes as well as the high plateaus known as the Altiplano and Puna. The slab in this segment can be traced to a depth of nearly 600 km beneath the foreland of the Andes. At 15°S beneath southern Peru, the subducted slab again flattens to a near horizontal attitude below 100 km depth. This northern "flat slab" underlies most of Peru. Like their southern counterparts, the high Andes here are quite narrow and recent volcanism is absent.

It has been proposed that subduction of the Nazca and Juan Fernandez oceanic ridges (Figure 22.3.2) are responsible for flat subduction. The latter does

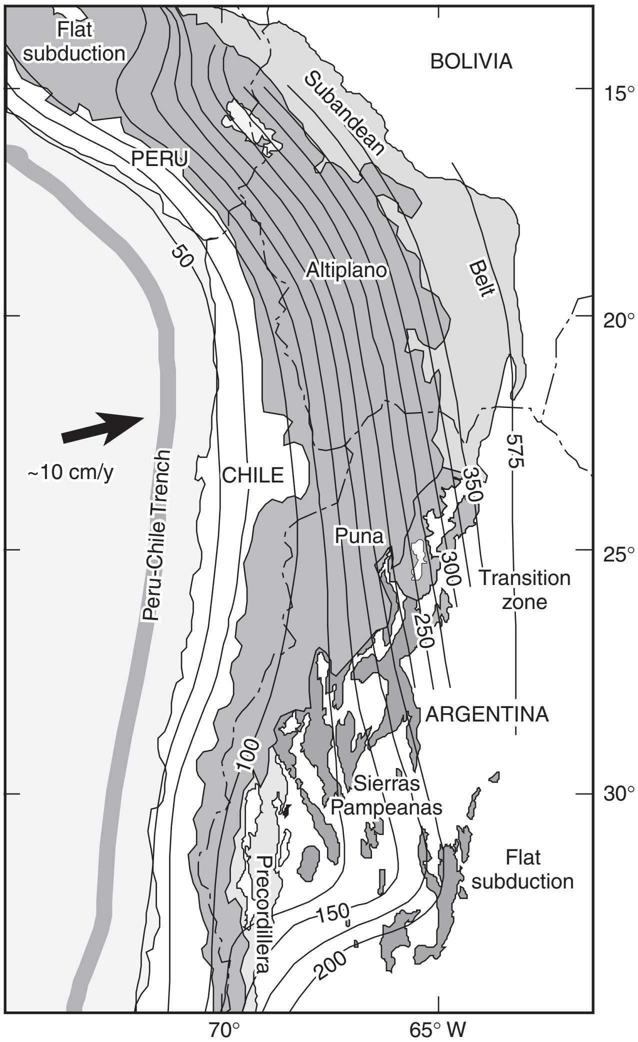

FIGURE 22.3.4 Tectonic provinces of the central Andes. Thin smooth lines are contours on the Wadati-Benioff zone (interval = 25 km). The medium gray region marks the area above 3 km elevation. Foreland thin-skinned thrust belts are shown in light gray; thick-skinned provinces in dark gray. Note that the thrust belt is not continuous but is intersected by the thick-skinned Sierras Pampeanas.

coincide closely with the boundary between flat and steep subduction at ~33°S, but the former is located 1° to 2° north of the flat-steep boundary beneath southern Peru. The plate kinematics clearly shows that the ridges have swept progressively southward along the plate boundary, but the continental geology shows no evidence of this effect.

Tectonic Segmentation of the Central Andes

The geology of the Late Cenozoic Andes to a first order reflects the lateral segmentation of the subducting plate (Figure 22.3.5). In our discussion, we will concentrate on two swaths across the Andes, one over the nearly flat segment of the subducted Nazca Plate between 28° and 33°S and the other over the 30°E-dipping segment between 15°S and 24°S, with brief references to other parts of the Andes.

The Andes from 15°S to 24°S. This segment of the central Andes overlying a 30°E-dipping subducted slab most closely approximates the average geologist's image of the Andean orogen "type" (Figures 22.3.4 and 22.3.5a). A cross section from west to east across this segment would show the plate boundary, a forearc region dominated by the longitudinal valley of northern Chile, the active volcanic arc, a high continental plateau system bounded on the east by the Eastern Cordillera, and the low-lying Subandean belt of thrusts and folds. The segment is dominated by the Altiplano-Puna Plateau, a region of more than 500,000 km^2 elevated to an average height of 3.7 km. There has been active volcanism in this segment for about the last 25 m.y. and, although this activity is focused in the Western Cordillera, locally it reaches the eastern edge of the plateau system. The distribution of volcanic rocks, modern morphology, ancient geomorphic surfaces, and structural geometry have led to the proposal of a two-stage uplift model for the plateau. During the Miocene, the crust beneath the current plateau was thermally weakened, resulting in horizontal shortening of the entire crust. At about 10 Ma, shortening mostly ceased in the plateau and in the Eastern Cordillera and migrated eastward into the Subandean belt, a thin-skinned foreland fold-thrust belt. At that point, the cold lithosphere of the Brazilian Shield began to be thrust beneath the mountain belt.

The Andes from 28°S to 33°S. This segment overlies the southern flat subduction zone (Figures 22.3.4 and 22.3.5b). The high Andes (above 3 km elevation) are narrow, but include Mt. Aconcagua, at 7 km the highest peak in the Western Hemisphere. Although they are largely composed of volcanic rock, they are not volcanic edifices but structural uplifts; magmatism has been lacking in this segment for the last 5–10 m.y. The magmatic history of the segment indicates that the subducted plate beneath Argentina shallowed between 16 and 6 Ma. There is some evidence from the flat subduction segment in Peru to indicate that shallowing of the slab began as recently as 2 Ma. The cessation of volcanism where subduction is flat is thought to be due to the virtual lack of asthenosphere between the subducted and overriding plates.

In the Argentine flat segment, the Sierras Pampeanas are thick-skinned basement uplifts with structural geometries very reminiscent of the Laramide Rocky Mountain foreland of the western United States (Figure 22.3.3). Like the Rocky Mountain foreland,

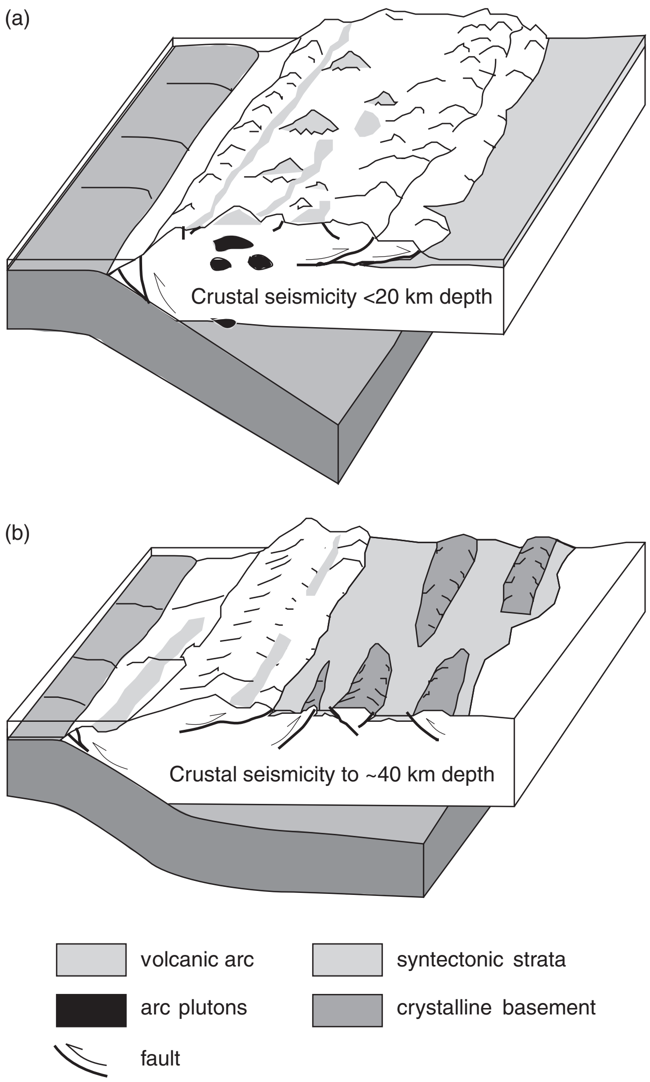

FIGURE 22.3.5 Schematic block diagrams showing the tectonic segmentation of the central Andes.

many blocks of the Sierras Pampeanas display an exhumed Paleozoic-Early Mesozoic erosional surface that has been rotated, probably by movement on listric faults, during the last 5 m.y. Seismicity associated with the westernmost of these basement uplifts indicates brittle failure, at least on short timescales, to depths of 40 km, nearly the entire thickness of the continental crust. This is unusually deep for continental deformation and is well below the predicted depth for the brittle–plastic transition, using even the most conservative parameters for the frictional and power law rheologies and heat flow on which those models are based. Because of extensive jungle cover, much less is known of the style of foreland deformation located over the Peruvian flat segment north of 15°S.

Although the thick-skinned basement uplifts are the most obvious features of the foreland overlying the Argentine flat slab, thin-skinned deformation also occurred between the Sierras Pampeanas and the High Andes in the Precordillera (Figure 22.3.4). Thrusting there began at about 20 Ma and has continued to present where it has not been buttressed by the Sierras Pampeanas. Thus, as in the western United States, thin- and thick-skinned thrusting overlapped in time, and both were active during flat subduction and during a period without significant magmatism. Shortening is about 5% in the Sierras Pampeanas and >60% in the Precordillera.

HORIZONTAL EXTENSION AND STRIKE-SLIP FAULTING Horizontal extension within convergent mountain belts has been of considerable interest since the 1980s. Young extensional deformation in the central Andes is concentrated at the northern and southern ends of the Altiplano-Puna Plateau. In the Puna, horizontal extension is mostly related to strike-slip and oblique-slip faults, whereas normal faults dominate at the northern end of the Altiplano. In both areas, however, horizontal extension is oriented approximately north-south, that is, subparallel to the strike of the orogen. There is little firm evidence for significant extension perpendicular to the belt, and the main part of the plateau system is neotectonically and seismically quiescent. There is probably significant strike-slip faulting in the Bolivian Eastern Cordillera north of the bend at ~18°S, which may accommodate the more oblique angle of convergence in this part of the belt. Even though extension is most notable around the ends of the high plateau, it is not restricted to the high topography. In northwest Argentina, horizontal extension occurs at elevations as low as 900 m in the foreland, and in southern Peru normal faults are found on the Pacific coast.

South of 28°S, several long fault zones, including the Liquiñe-Ofqui Fault in southern Chile and the El Tigre Fault in western Argentina, have undergone significant strike-slip movement during the Quaternary. These faults are thought to be due to the slight nonorthogonal convergence between the Nazca and South American Plates. The Atacama fault system of northern Chile is probably best known for its strike-slip history, but recent work has shown that it has a much more complicated and protracted history, including dip-slip and strike-slip displacements during the Mesozoic to Middle Tertiary and mostly dip-slip motion since then.

OROCLINAL BENDING The marked curvature of the Andes at about 18°S raises the question of whether this curvature reflects the initial shape of the continental margin, or whether it is a product of Andean

TABLE 22.3.1 THRUST BELT CHARACTERISTICS, BOLIVIA AND ARGENTINA

Location	Width (km)	Shortening (km)	Topographic Slope	Wedge Taper	Annual Precipitation (mm)
N. Bolivia 13°–17°S	70	115 ± 20	3.5°	7° ± 1°	1000–2800
S. Bolivia 19°–23°S	90–110	75 ± 10	0.5–1.0°	2.5° ± 1°	400–100
Precordillera Argentina 29°–33°S	40–60	105 ± 20	~2.5°	3.5° ± 1.5°	100–200

deformation. The answer is probably both. Paleomagnetic data derived from Mesozoic rocks along the west coast of South America show clockwise rotations south of the bend and counterclockwise rotations to the north. The main debate over these results concerns whether the rotations reflect in situ block rotations or regional oroclinal bending. To date, several carefully mapped sites in the foreland provide evidence only for local, rather than regional, rotations. These preliminary results, however, do not preclude the model we have presented, in which the curvature of the central Andes has been accentuated during the last 25 m.y. by laterally variable shortening that is greatest within, and on the margins of, the Altiplano-Puna Plateau, and decreases to both north and south.

PALEOTECTONIC CONTROL Although structural style within the central Andes shows a broad correlation with the geometry of the subducted plate, as we have described, preexisting heterogeneities within the continental crust also play an important role. Thin-skinned thrust belts are restricted to thick, wedge-shaped Paleozoic basins. The Subandean belt is largely located within a previously undeformed Paleozoic passive margin and foreland basin sequence, east of a zone of preexisting deformation now occupied by the Eastern Cordillera. The Precordillera thrust belt deforms a lower Paleozoic passive margin sequence of what is called the Precordillera terrane, a narrow slice within western Argentina that bears marked similarities to the Lower Paleozoic of the southern Appalachians. The transition from the Subandean belt to the northern Sierras Pampeanas thick-skinned deformation coincides with southward thinning of Paleozoic strata, and the western boundary of the Sierras Pampeanas is the boundary of the Precordillera Terrane. Finally, many complex local structures within the Bolivian Altiplano and in northwestern Argentina owe their geometries to reactivation of Late Cretaceous rift basins.

22.3.4 Crustal Thickening and Lithospheric Thinning

Modern estimates suggest that magmatism has contributed less than 10% to the total crustal thickening in the Andes during the last 25 m.y. Thus, the rest of the topography in the Andes must be accounted for by two mechanisms: (1) thickening of the crust by deformation, and (2) thermally controlled thinning of the lithosphere giving rise to uplift. Most of the crustal shortening responsible for the present topography is manifested at the surface as the thin-skinned thrust belts that provide the interface between the Andes and the Brazilian Shield. The thrust belts of the central Andes, including the Subandean belt and the Precordillera, are of considerable interest because they are among the few active examples of an antithetic (or foreland-verging) foreland thrust belt, in which the overall sense of shear is opposite to that in the associated subduction zone. The along-strike variations in these thrust belts allow one to identify the key first-order associations of geometry, topography, shortening, and paleotectonic setting (Table 22.3.1). There is a general correlation among high critical wedge taper, width, and a high-degree of shortening. In contrast, the order-of-magnitude variation in precipitation (and, presumably, erosion rate) shows no clear effect on the shortening.

Crustal thickening alone does not appear to be sufficient to explain the high plateau (the Altiplano) of the central Andes. About 1 to 1.5 km is probably accounted for by thinning of the lithosphere beneath the plateau. When this thinning occurred and how it relates to crustal shortening remain unresolved problems. As has been pointed out for the Alpine system, horizontal shortening should thicken not only the continental crust, but also the entire lithosphere. Yet beneath the plateau, the continental lithosphere must have thinned even as the crust was thickening. Furthermore, in the regions of flat subduction, the subducted plate is only 100–120 km beneath the surface, even though "normal" continental lithosphere is thought to be on the order of 150 km thick. Because it takes many millions of years to thin the lithosphere by conduction alone, recent proposals have invoked delamination of the base of the lithosphere to produce the necessary thinning on the timescale implied by Late Cenozoic mountain building in the Andes.

22.3.5 Closing Remarks

The Andes present a unique natural laboratory for studying mountain building that has occurred without the aid of a collision between two continental masses. The most important processes that have produced the modern topography of the Andes are structural shortening and lithospheric thinning. Volcanism, in contrast, is responsible mostly for the volumetrically minor topography above 4 to 4.5 km. Magmatism, nonetheless, probably plays a very significant role in determining the rheology of the crust, producing weak zones subject to faulting.

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22.4 THE APPALACHIAN OROGEN—An essay by James P. Hibbard[1]

22.4.1 Introduction

The **Appalachian Orogen** (Figure 22.4.1) is a northeast-trending belt of Late Precambrian to Paleozoic rocks in eastern North America that were deformed during the Paleozoic. Its strike length is more than 3000 km, extending from Alabama to Newfoundland, and it forms a segment of a much larger Paleozoic orogenic system that encompasses the Caledonide Orogen of the British Isles, Greenland, and Scandinavia to the northeast, and the Ouachita Orogen to the southwest. The northwest limit of the Appalachians is the deformation front between rocks of the orogen and older orogens and platform rocks of North America. On its southern and southeastern flanks, the orogen is onlapped by Cenozoic sedimentary rocks of the Atlantic Coastal Plain.

The orogen is important from a historical standpoint, as many significant tectonic ideas are rooted in Appalachian bedrock. The concept of geosynclines, which dominated thought on mountain building for more than a century preceding the advent of plate tectonics, was conceived in the Appalachians. Appropriately, the geosynclinal theory was supplanted on its native Appalachian turf by plate tectonics in the 1960s and 1970s. Initial questioning of whether the Atlantic closed and then reopened was followed by the first detailed application of plate tectonic theory to an ancient orogen. In addition, the idea of "thin-skinned tectonics," meaning the deformation of cover strata above a master décollement that is independent of underlying basement, was first developed in the classic Valley and Ridge fold-and-thrust province of the southern Appalachians. Closely related to the thin-skinned concept was the realization that there is a mid-crustal detachment within the orogen that places a large portion of the deformed southern Appalachians onto the relatively undeformed North American platform. This realization lead to the general acceptance that substantial portions of orogenic belts form relatively thin, highly allochthonous sheets (or tectonic flakes) emplaced onto cratons.

Having established the prominent role of the Appalachians at the forefront of tectonic research, we turn, in the remainder of this essay, to a sketch of current thought on the orogen. Following an overview of the mountain belt, there is a brief description of the first-order crustal components, an explanation of how and when they were assembled, and finally a highlighting of potential directions for future thought and development of Appalachian tectonics.

22.4.2 Overview

The structural grain of the Appalachians is remarkably consistent, defining a series of broad, harmonically curved promontories and reentrants (Figure 22.4.1); their grace and regularity lead the renowned North American tectonicist, P. B. King, to proclaim the Appalachian Orogen as the most elegant mountain chain on earth. As we will see, this structural architecture reflects a fundamental feature of the orogen that was important in its evolution. For our purposes, the New York promontory will serve as the divide between segments referred to as the northern and southern Appalachians.

In contrast to structural divisions, lithotectonic divisions of an orogen distinguish rock associations that were either formed or deposited in a common tectonic setting during a finite time-span. These divisions are scale dependent, that is, contingent on the scale of the tectonic process considered. In this essay, the hierarchy of lithotectonic divisions consists of the realm at orogen scale and the zone at the scale of two or less reen-

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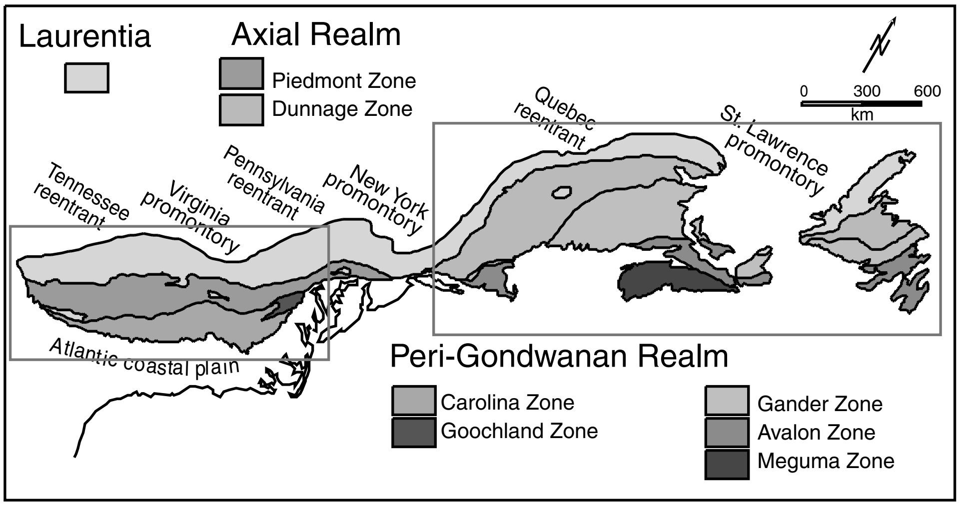

FIGURE 22.4.1 Realms and zones of the Appalachian Orogen, defined on the basis of Middle Ordovician and older geologic history. The boxes outline areas shown in Figures 22.4.4 and 22.4.5.

trants. At yet smaller scale, terranes are recognized as regional subdivisions of a zone; however, in this essay we will mainly be concerned with realms and zones.

The Appalachians are composed of three realms, the Laurentian realm,[2] the Axial realm, and the peri-Gondwanan realm (Figure 22.4.1), all of which acquired their defining geologic character before the Middle Ordovician. The Laurentian realm encompasses essentially all of the rocks deposited either on, or adjacent to, ancient North America and forms the western flank of the orogen; however, windows of Laurentian rocks occur locally among the more easterly accreted terranes (Figure 22.4.1). In contrast, peri-Gondwanan realm rocks are interpreted to have formed proximal to Gondwana[3] and thus are considered to be exotic with respect to Laurentian elements; they are distributed along the eastern flank of the Appalachians. The Axial realm is a collection of zones and terranes of mainly oceanic and volcanic arc affinity that has been caught between the Laurentian and peri-Gondwanan realms during Appalachian orogenesis. Unlike the uniformity of the Laurentian realm, both the peri-Gondwana and the Axial realms change character along strike of the orogen.

The orogen was assembled during the approximately 300 m.y. time-span between the existence of two supercontinents, the Middle Proterozoic Rodinia and the Late Paleozoic Pangea. The Appalachians formed as a result of the progressive accretion of Axial and peri-Gondwanan elements to the Laurentian realm. Classically, it has been accepted that three major events, the Taconic, Acadian, and Alleghanian Orogenies record the accretion of these elements. However, as more data are acquired, we are finding that accretionary events along the Laurentian margin were continuous for protracted periods of time and were less punctuated than is implied by the simple mantra of “Taconic, Acadian, and Alleghanian.”

22.4.3 Tectonic Components

THE LAURENTIAN REALM The template for Appalachian accretionary events, the eastern Laurentian continental margin, was initiated by Late Precambrian rifting along the axis of the ~1-Ga Grenville Orogen within Rodinia; thus Grenville rocks form basement to the continental margin. The west flank of the Amazonian craton likely formed the conjugate margin to eastern Laurentia during this extensional event. The main pulse of rifting affected the entire length of the Appalachians at roughly 600–550 Ma. Sedimentation was synchronous with rifting, leading to thick deposits confined to elongate basins and characterized by abrupt changes in the thickness of strata, with most of the sediment supplied from the Laurentian craton. Bimodal magmatism accompanied rifting; however, there are two pulses of rift magmatism, an early pulse (~750–700 Ma) (Figure 22.4.2) confined to the southern Appalachians, and a later, main pulse (~600–550 Ma) along the length of the orogen. The

[2]Laurentia refers to early Paleozoic North America, including portions of Greenland, Scotland, and Ireland that rifted away in the Mesozoic.

[3]Gondwana is the ancient continent approximately equivalent to an amalgamation of modern southern hemisphere cratons.

early pulse appears to be coeval with the early breakup of Rodinia along the Pacific margin of Laurentia and may well be a far-field effect of this event.

Rifting led to continental breakup, the onset of seafloor spreading, thermal subsidence of the newly formed passive margin, and deposition of a drift sequence atop the rift deposits (Figure 22.4.2 and 22.4.3). The drift sequence consists of basal clastic rocks overlain by a shallow marine carbonate sequence up to 10 km thick. In contrast to the heterogeneity of the rift deposits, the drift sequence displays an orderly stratigraphy with little thickness variation and remarkable lateral continuity. Paleomagnetic studies indicate that the margin was at near equatorial latitudes during establishment of the passive margin. The seaward edge of this extensive carbonate shelf is marked by a facies change into deep water shaley rocks that locally contain carbonate blocks and boulders that spalled off the precarious edge of the shelf. The Paleozoic ocean that formed outboard of eastern Laurentia, preceding the modern Atlantic, is called Iapetus, after the mythical Greek father of Atlantis.

The geometry of the continental margin was controlled by the zigzag pattern formed by spreading and transform segments of the rift system. This shape influenced the distribution of rift and drift sequences; former ridge-transform junctions along the margin tended to form steep-sided terminations for rift basins, whereas the distribution of the drift sequence facies change from shelf to slope-and-rise was controlled by the jagged shape of the margin. Where the trace of the drift sequence facies change is preserved, it presently follows the curves of the structural promontories and reentrants in the orogen. This relationship indicates that the promontories and reentrants are inherited from the original shape of the margin at breakup.

The Axial Realm

Elements of the Axial realm record the evolution of Iapetus and its component volcanic arcs, backarc basins, and accretionary complexes. There appears to be a major change in the realm at the New York promontory; in the northern Appalachians, the Dunnage Zone records a complex history of multiple volcanic arcs and backarc basins, whereas in the southern Appalachians, the Piedmont Zone appears to record a simpler history of a single composite arc system.

The Dunnage zone is in tectonic contact with Laurentian rocks along the Baie Verte–Brompton line (Figure 22.4.4), which is a steep, relatively narrow fault system that has experienced multiple episodes of movement; in many places it is marked by narrow, elongate ultramafic bodies. The zone is best exposed along the north coast of Newfoundland, where its entire width is at low metamorphic grade; here, the zone records the evolution of at least two distinct oceanic tracts. Contrasts in stratigraphy, fossil faunas, and paleomagnetic and isotopic data indicate that the western tract of the zone was associated with the Laurentian side of Iapetus, whereas the eastern tract developed on the Gondwanan side of the ocean realm. They are tectonically juxtaposed along the Red Indian Line. Both tracts record an early arc phase that starts in the Early to Middle Cambrian and a younger arc phase that ranges into the Late Ordovician (Figure 22.4.3b).

Elements of the Newfoundland Dunnage zone can be correlated with units in New Brunswick, Quebec, and northern Maine; however, most Dunnage zone elements from central Maine to New York are multiply deformed and have been subjected to high-grade metamorphism, thus obscuring original relationships between units. Consequently, the early evolution of the zone is not as well understood in New England, although there are strong hints that it conforms to that of the Canadian Dunnage zone.

The Piedmont zone is tectonically severed from Laurentian rocks to the west along a series of faults, most of which have multiple movement histories. Much of the zone has been subjected to intense, polyphase deformation and medium- to high-grade metamorphism and thus unraveling the depositional-magmatic history is somewhat tenuous. The zone is split into two components by the Brevard zone, a polygenetic ductile shear zone, and other faults northward along strike of the Brevard zone (Figure 22.4.5).

The western portion of the zone is dominated by metamorphosed clastic rocks and associated mélanges disposed in imbricate thrust stacks. In northern Virginia, the thrust stacks were assembled by the Early Ordovician, but locally, in southwest Virginia, radiometric ages suggest that the clastic rocks have been subjected to Early Cambrian shortening and medium-grade metamorphism (Figure 22.4.2). In North Carolina, metamorphosed clastic rocks at the western edge of the zone contain pods of eclogite. Across the Brevard zone, the eastern portion of the zone contains substantially more metamagmatic rocks than the western area. Magmatism appears to have been active from the Early Ordovician to the Early Silurian, with a Late Ordovician lull during which black slates were deposited in central Virginia (Figure 22.4.2). Where studied, the magmatic rocks are geochemically consistent with a suprasubduction zone, volcanic arc setting.

Despite the strong tectonothermal overprint, all of the characteristics just outlined are consistent with the

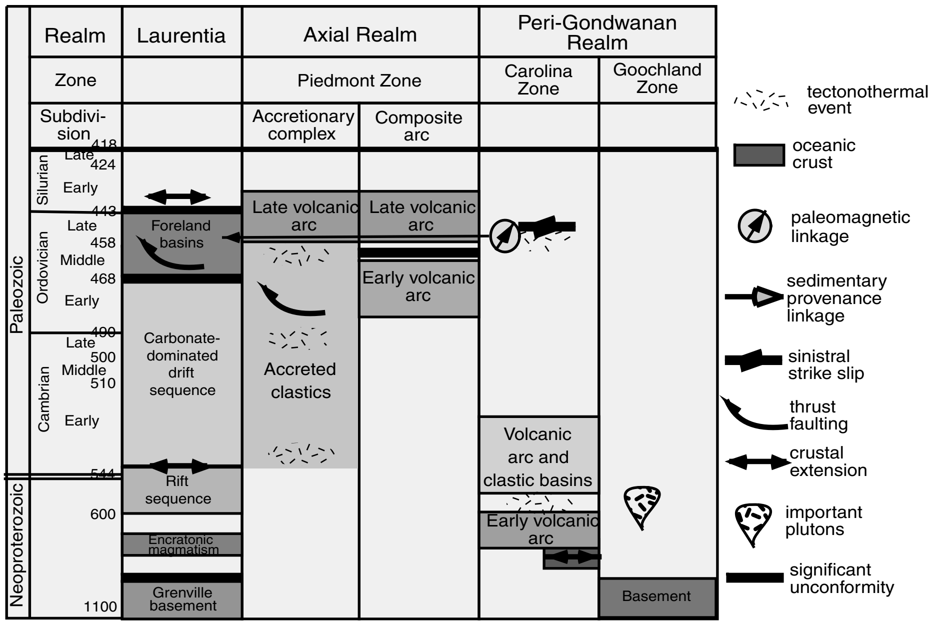

FIGURE 22.4.2 Major elements and events of the southern Appalachians from the Precambrian to the Silurian.

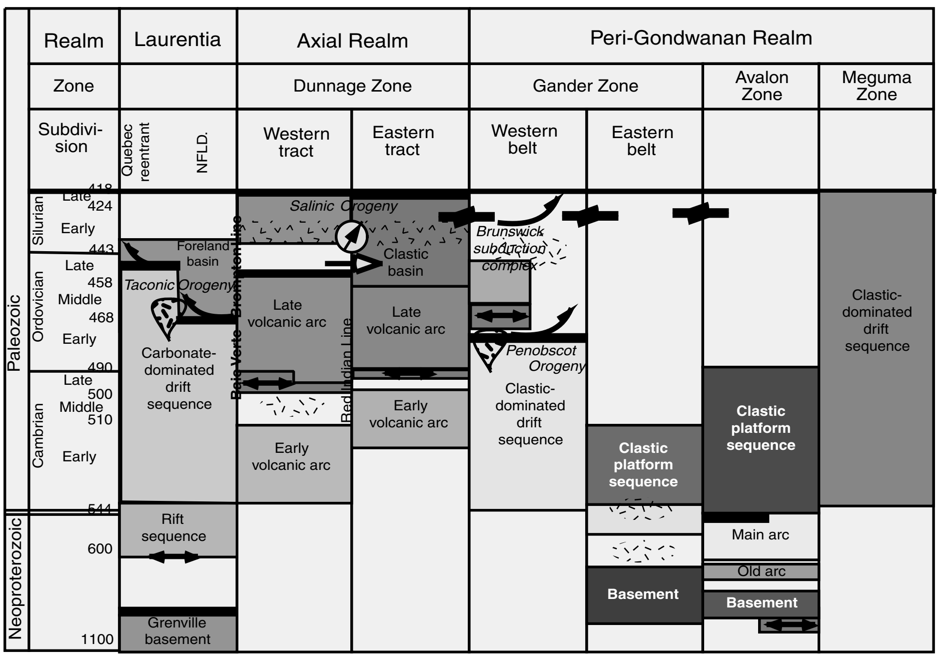

FIGURE 22.4.3 Major elements and events of the northern Appalachians from the Precambrian to the Silurian. See Figure 22.4.2 for explanation of symbols.

interpretation that the Piedmont zone encompasses a long-lived, west-facing accretionary complex in front of a more easterly suprasubduction-zone magmatic arc. If the interpretation of an Early Cambrian tectonothermal event is valid, it suggests that the Piedmont Zone formed in an ocean older than Iapetus, which was just in its rift-to-drift stage at this time.

The Peri-Gondwanan Realm

A collection of diverse crustal remnants of Proterozoic to Early Paleozoic rocks that lay across the Iapetus Ocean from the eastern Laurentian margin is grouped here as peri-Gondwanan elements. In the northern Appalachians, the realm is represented by the Gander, Avalon, and Meguma zones, whereas the Carolina and Goochland zones occupy the east flank of the exposed southern Appalachians (Figures 22.4.1, 22.4.4, and 22.4.5). The nature of the Goochland zone is controversial and it is tentatively grouped here with the peri-Gondwanan Zones.

The Gander zone is in both fault and stratigraphic contact with the Dunnage zone. For example, in central Newfoundland, the contact is marked by a thrust fault that emplaces Dunnage ophiolite on top of the Gander zone, whereas in northern Maine and New

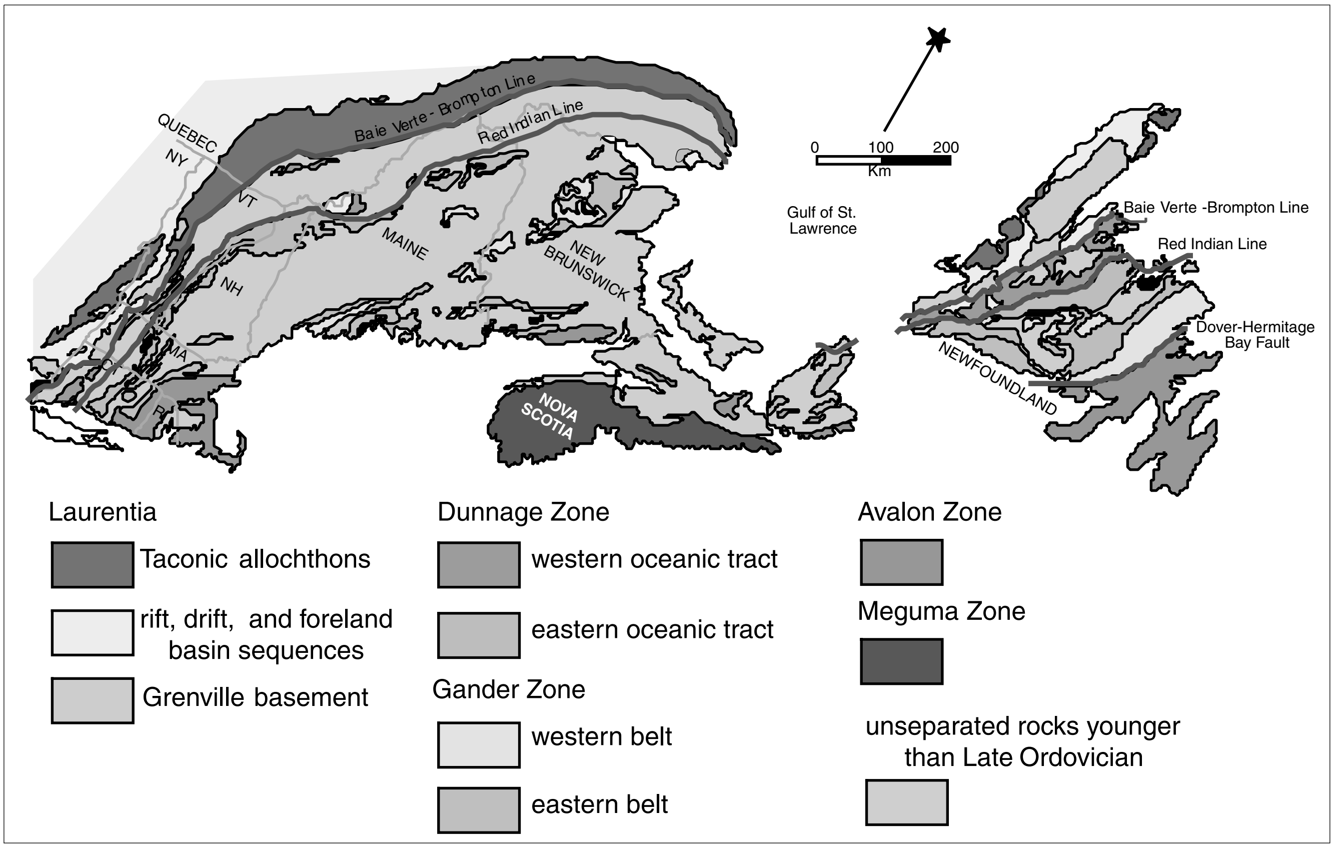

FIGURE 22.4.4 Distribution of Middle Ordovician and older elements in the northern Appalachians.

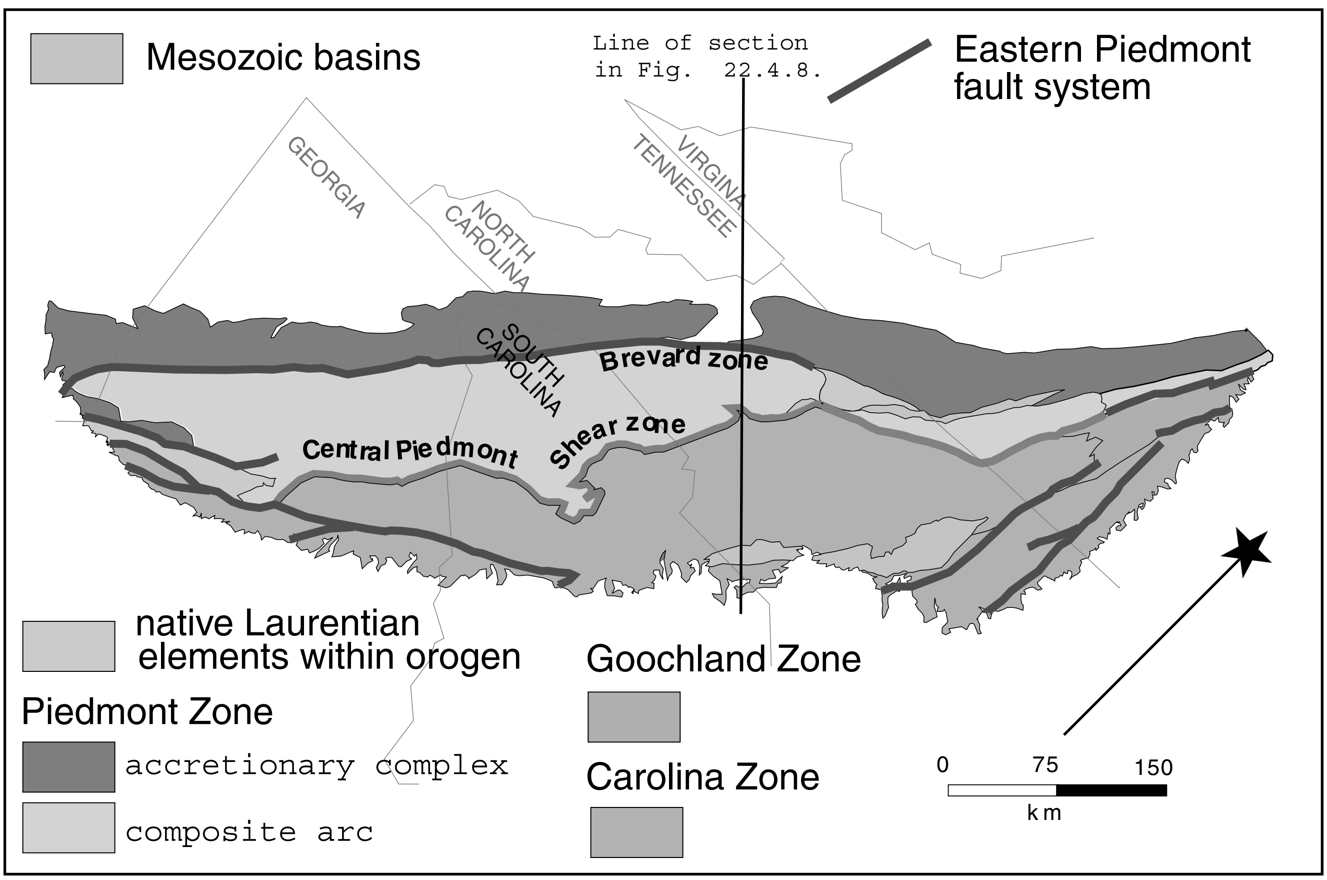

FIGURE 22.4.5 Distribution of Middle Ordovician and older elements in the southern Appalachians. Line of section is for Figure 22.4.8.

Brunswick, Dunnage volcanic rocks unconformably overlie the Gander zone (Figure 22.4.4). Elements grouped here as the Gander zone define two distinct belts on either side of the zone (Figures 22.4.3, and 22.4.4). The eastern belt (called the Avalon zone by some workers) is composed of older crystalline basement that is faulted against a younger magmatic sequence. Basement rocks consist of marble, quartzite, greenstone, and pelite that appear to have protolith ages greater than 800 Ma and that have been involved in metamorphic events prior to 600 Ma. The younger magmatic sequence includes mafic and felsic volcanic and plutonic rocks with an approximate age range of 600–545 Ma. The western belt is characterized by Cambrian-Early Ordovician continentally derived quartz arenite and pelite with minor mafic magmatic rocks that, collectively, have been interpreted as representing a passive margin sequence. The eastern belt may form the basement on which the western belt passive margin was deposited, but the contact between the two is unknown.

The Avalon zone is in fault contact with the Gander zone; in Newfoundland the contact is represented by the Dover-Hermitage Bay Fault (Figure 22.4.4), which is vertical; it is documented on seismic reflection profiles as reaching the base of the crust. The zone is dominated by Neoproterozoic magmatic rocks that exhibit diverse compositions and have mainly suprasubduction zone signatures. Geochemical studies indicate that the volcanic pile likely formed on thin continental crust, although this basement does not appear to be exposed. Magmatism extended over the broad time period of 685–540 Ma (Figure 22.4.3), with a preponderance of activity in the range of 630–580 Ma. Locally, deformation was synchronous with deposition and appears to have been dominated by extension.

The magmatic rocks are overlain by a Lower Paleozoic transgressive, shallow marine platform sequence (Figure 22.4.3); the lack of substantial carbonate in this sequence as well as paleomagnetic data attests to the platform's being deposited at high paleolatitudes. Fossil faunas in the platform sequence are of "Avalonian" affinity, distinct from those of the Laurentian and Axial realms.

The Meguma zone underlies most of southern Nova Scotia and forms the southeastern most exposed crustal block in the orogen (Figures 22.4.1 and 22.4.4). It is faulted against Carboniferous cover rocks to the north, which in turn are unconformable upon the Avalon zone. The zone is dominated by Early Paleozoic turbidites that have been interpreted as being deposited in an abyssal fan setting along a passive continental margin. On the basis of sedimentology, stratigraphy, paleontology, petrology, and geophysics, the zone has been correlated with rocks in Morocco.

The Carolina zone is in tectonic contact with the Piedmont zone along the central Piedmont shear zone, a Late Paleozoic thrust fault (Figure 22.4.5). The zone is an amalgamation of Neoproterozoic to Early Paleozoic volcanic arcs and associated sedimentary rocks that have an approximate age range of 675–530 Ma (Figure 22.4.2). It appears that one or more deformational events coincided with magmatism, although the nature of these events is poorly known. The Carolina zone resembles the Avalon zone, but appears to be distinct from its northern Appalachian counterpart. Neoproterozoic magmatism in the Carolina Terrane peaked at 630–610 Ma and again at ~550 Ma, whereas in Avalon, peak magmatism is in the period 630–580 Ma. Although the Carolina zone contains an Early Paleozoic clastic sequence, it does not appear to represent a transgressive platformal sequence as found in Avalon. Also, Carolina fossil faunas have a peri-Gondwanan affinity, but are not "Avalonian."

The Goochland zone is in tectonic contact with the Piedmont zone and is likely faulted against the Carolina zone (Figure 22.4.5). The zone comprises orthogneiss and paragneiss that have been intruded by anorthosite dated at ~1 Ga. In addition, this package is intruded by alkalic granite dated at ~630 Ma. The zone may represent a structural window into Laurentian basement, with the younger granite representing rift magmatism; however, the 630-Ma age of this granitoid does not coincide with known rift magmatism on the Laurentian margin. Alternatively, the zone represents peri-Gondwanan basement with the granitoid equivalent with Neoproterozoic plutons in the Carolina zone.

22.4.4 Assembly

EARLY PALEOZOIC DESTRUCTION OF IAPETAN PASSIVE MARGINS The eastern Laurentian passive margin came to an abrupt demise in the Early to Middle Ordovician. This event, termed the Taconic Orogeny, is marked by a regional unconformity on the continental shelf, a change in sedimentation along the margin, the development of a submarine thrust belt, and accompanying metamorphism; it is best preserved in the northern Appalachians (Figure 22.4.3).

Carbonate sedimentation along the Laurentian shelf, slope, and rise was choked off in the Early to Middle Ordovician, and a Middle Ordovician erosional to slightly angular unconformity was developed along the length of the carbonate shelf (Figure 22.4.3). The new sedimentary regime was marked by deep water, foreland deposition of easterly derived clastic sediments. The continental shelf and clastic foreland basin were overridden by thrust sheets containing Laurentian slope and rise- and rift-related sediments; the highest thrust sheets in Newfoundland and Quebec are composed of ophiolite. Where they are emplaced upon the passive margin, there is a striking contrast between autochthonous rocks of the shelf and allochthonous, deeper water, rocks of the thrust sheets; the conspicuous thrust sheets are termed the "Taconic allochthons." Perhaps one of the most inspirational geologic sights in the Appalachians is the view of the barren, flat-topped ophiolite sheet in eastern Newfoundland (now a UNESCO World Heritage Site).

Taconic events are interpreted as reflecting the introduction of the Laurentian margin into a subduction zone beneath the eastern tract of the Dunnage zone (Figure 22.4.6a). The unconformity represents the flexural bulge due to loading of the continental margin by an overriding accretionary complex, the clastic sedimentation represents foreland basin, or trench, sedimentation on top of the downgoing continental margin, and the thrust sheets represent an accretionary wedge. However, this tectonic system must have been more complex than the simple subduction of Laurentia beneath the Dunnage zone, for some of the obducted ophiolites were just forming while obduction was in progress elsewhere along the margin. Additionally, Early Ordovician plagiogranites intrude obducted ophiolites; in contrast to the obduction process, these plutons and other geologic evidence require a rapid change in subduction polarity involving west-directed subduction beneath the margin and volcanic arc (Figure 22.4.6b).

The Taconic Orogeny in the southern Appalachians appears to represent the attempted subduction of the Laurentian margin beneath the Piedmont zone accretionary prism and arc. However, the foreland clastic wedge in the southern Appalachians is overlain by that of the northern Appalachians, indicating that the Taconic event was slightly older in the south. Also, the hallmark Taconic thrusting and metamorphism have been severely overprinted and obscured by younger tectonothermal events.

Nearly synchronous with the Taconic Orogeny along the Laurentian margin, the Gander passive margin of Iapetus was also tectonically terminated. This event, the Penobscot Orogeny, is recognized by an unconformity of Early Ordovician volcaniclastic rocks affiliated with the Dunnage Zone atop Cambrian-Early Ordovician Gander quartzose clastic rocks. This unconformity persists along strike from northern Maine to Newfoundland. In Newfoundland, the Penobscot Orogeny also involved the eastward obduc-

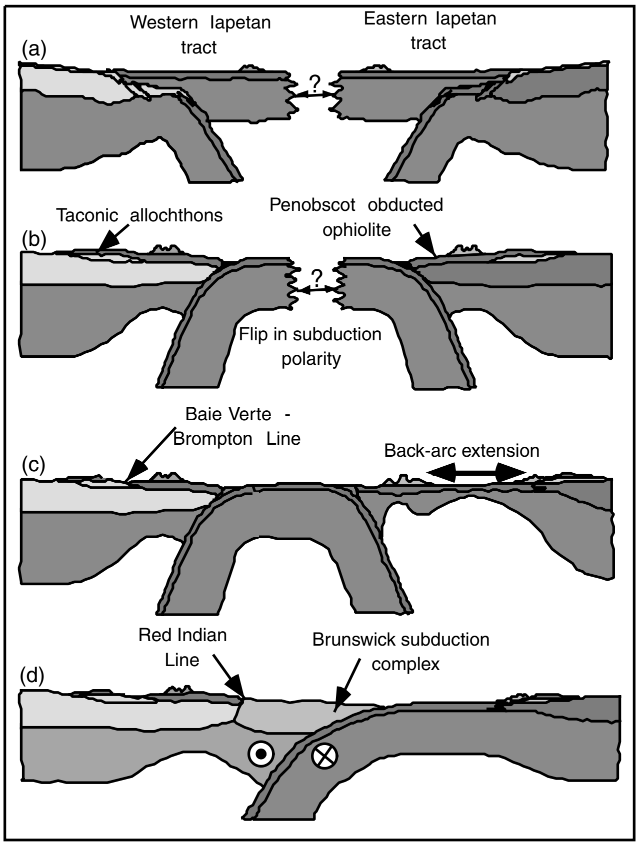

FIGURE 22.4.6 Cartoon depicting possible tectonic evolution of the northern Appalachians: (a, b) Early to Middle Ordovician; (c, d) Middle to Late Ordovician.

tion of Dunnage zone ophiolite onto the Gander passive margin (Figure 22.4.6b). The timing of Penobscot obduction is tightly constrained by Early Ordovician fossils in the overthrust oceanic rocks and by an Early Ordovician granitoid that intrudes and "stitches" both Gander and Dunnage zone rocks. Thus, Iapetus commenced closure from both margins in the Early Ordovician.

MID-PALEOZOIC CLOSURE OF IAPETUS Late Ordovician to Late Devonian events that contributed towards the closure of Iapetus and construction of the Appalachians are best recorded in the northern Appalachians, where Silurian and younger strata blanket the orogen. In the southern Appalachians, Middle Paleozoic strata are largely confined to covering the Laurentian realm. However, the earliest interaction of Laurentia with the peri-Gondwanan realm is apparently recorded in the southern Appalachians, where circumstantial evidence from across the orogen indicates that the Carolina zone commenced docking in the Middle to Late Ordovician, immediately on the "coattail" of the Taconic accretion of the Piedmont zone (Figure 22.4.2).

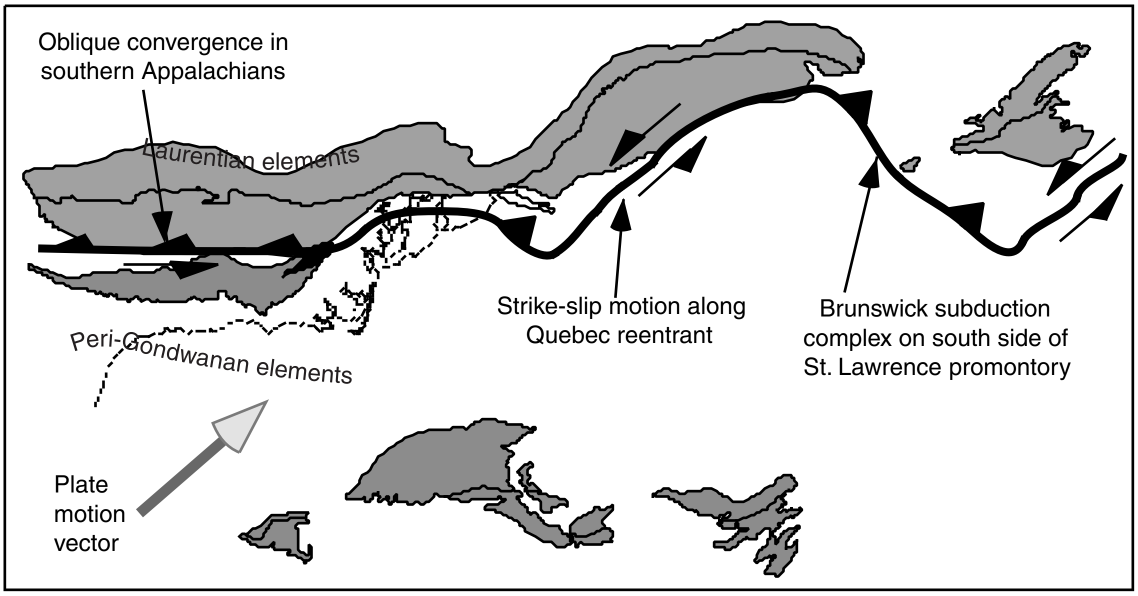

FIGURE 22.4.7 Cartoon of possible tectonic setting during the Late Ordovician to Silurian Salinic Orogeny. The interaction between Laurentia and peri-Gondwana elements involved a strong component of sinistral shear, which may have produced oblique subduction in the southern Appalachians, and sinistral shear and convergence along reentrants and promontories, respectively, of the northern Appalachians. Bold arrow shows approximate plate motion vector for peri-Gondwana elements relative to Laurentia.

In native Laurentian rocks of the southern Appalachians, an unconformity at the Ordovician-Silurian boundary has been attributed to tectonic loading of the post-Taconic margin (Figure 22.4.2). Furthermore, the post-Taconic margin of Laurentia, the Piedmont zone, was intruded by a pulse of Late Ordovician granodioritic to tonalitic plutons that likely reflects subduction beneath the Laurentian margin (Figure 22.4.2). Finally, in the Carolina zone, Late Ordovician upright folding, greenschist facies metamorphism, and uplift probably mark the initiation of the collision of Carolina with Laurentia, as paleomagnetic data indicate that it was at Laurentian paleolatitudes by this time (Figure 22.4.2). Folds in the Carolina zone define an *en echelon* array that is consistent with a component of sinistral shear during collision.

In the northern Appalachians, closure of Iapetus continued after the Late Ordovician flip in subduction polarity (Figure 22.4.6b), but it was counteracted by the generation of a backarc basin on the peri-Gondwanan side (Figure 22.4.6c). However, the western and eastern oceanic tracts of the Dunnage zone, which had distinct Early Paleozoic faunas, shared a mixed Late Ordovician fauna, indicating that they were proximal to one another. Additionally, similar Silurian paleomagnetic data from each tract as well as an Early Silurian stitching pluton along the trace of the Red Indian Line further support a Late Ordovician juxtapositioning of the two tracts (Figures 22.4.3 and 22.4.6c).

Evidence for the convergent closure of the Middle to Late Ordovician peri-Gondwanan backarc basin is preserved only along the east-west trending portions of the orogen in the vicinity of the St. Lawrence promontory; there, the Brunswick subduction complex, a southeast vergent stack of thrust sheets that includes Late Ordovician to Silurian blueschist and ophiolitic mélange, records this closure. Elsewhere, along more northeast-trending segments of the northern Appalachians, Silurian sinistral shear is recorded, from the Avalon zone across to the Laurentian margin (Figure 22.4.7). Thus, it appears that by the end of the Silurian, most components along the length of the orogen had been assembled along the Laurentian margin through the closure of Iapetus with a strong component of sinistral shear displacement (Figure 22.4.7); this kinematic regime, with regional shortening oriented approximately north-south, has been termed the Salinic Orogeny in the northern Appalachians.

At the end of the Silurian, the Laurentian margin underwent an abrupt change in kinematic character. Following a Late Silurian unconformity found in many places in the northern Appalachians, regional shortening reoriented to a position more at right angles to the trend of the orogen, with a component of dextral strike slip; this new kinematic regime is responsible for the

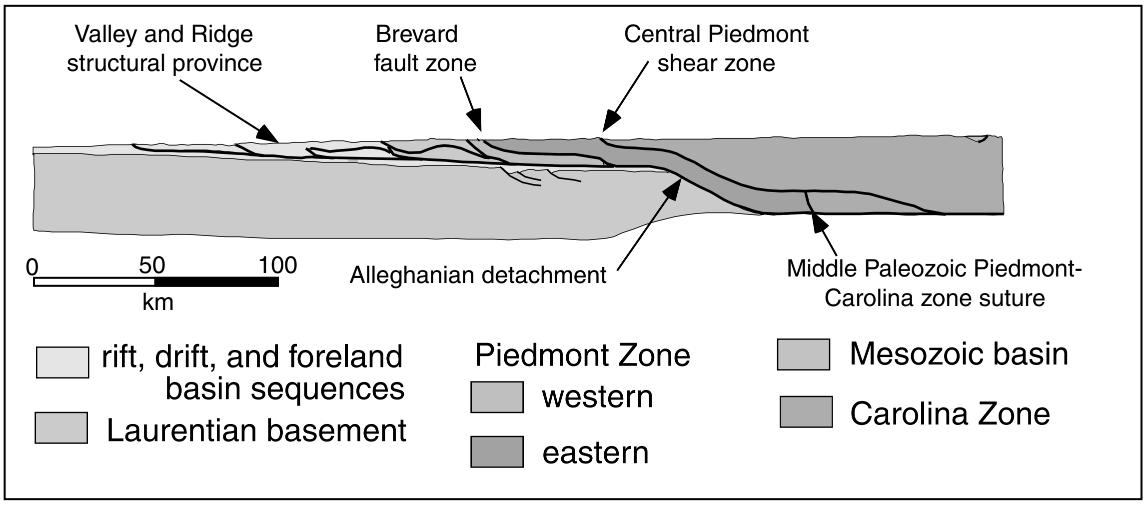

FIGURE 22.4.8 Cross section of the southern Appalachians showing major features of the Alleghanian thrust system. Line of section shown in Figure 22.4.6; no vertical exaggeration.

Acadian Orogeny. The change in kinematics is heralded by an Early Devonian westward transgressive clastic wedge in northern New England. However, the most intense manifestation of the Acadian event is recorded on the New York promontory, in southern New England, where Middle Paleozoic and older rocks were deformed into regional-scale recumbent fold nappes at high metamorphic grade. Subsequent rapid uplift resulted in the removal of up to 20 km of crust in the southern New England area and the deposition of the thick Devonian Catskill clastic wedge to the west. The plate-scale process responsible for the Acadian Orogeny may have been the docking of the Meguma zone to the orogen, for the Acadian is the first deformation shared by the Meguma rocks and the remainder of the orogen. Regardless, it is noteworthy that at the scale of the northern orogen, intense Acadian tectonism and uplift appear to be limited to the region of the New York promontory.

Late Paleozoic Formation of Pangea

The final phases of Appalachian orogenesis took place from the Mississippian to the Permian; events within this time frame are ascribed to the Alleghanian Orogeny. The orogeny is penetratively developed in the southern Appalachians as well as in southern New England; in both areas Alleghanian events strongly overprint earlier deformation and metamorphism; in contrast, it is more limited in development in most of the northern Appalachians.

In the southern Appalachians, the Valley and Ridge Province along the western flank of the orogen records the Late Paleozoic westward-directed thrusting of Cambrian to Permian strata onto the Laurentian platform. This thrust belt was studied in detail for more than a century before it was discovered that it represented merely the toe of what is now recognized as an orogen-scale thrust wedge (Figure 22.4.8). The full magnitude of this thrust wedge was only realized in a seismic reflection profile study across the southern orogen. It revealed that the Laurentian platform sequence extends in the subsurface to at least as far east as the central Piedmont shear zone. The crystalline thrust sheet encompasses the Piedmont zone and is separated from the underlying Laurentian platform by a major detachment fault; this geometry resolves into at least 175 km of shortening along the Alleghanian detachment.

The oldest documented west-directed thrusting is within the crystalline sheet along the central Piedmont shear zone in northern North Carolina; here Middle Mississippian granitoids are syntectonic with respect to ductile thrusting. This early thrusting is roughly coeval with the initiation of Carboniferous clastic wedges shed out over the Laurentian platform. Although thrusting continued into the Permian in the Valley and Ridge, there was a major change in kinematics in the Early Pennsylvanian in the eastern Piedmont and Carolina zones. There, thrusting was replaced by dextral strike-slip motion along a network of large faults, termed the eastern Piedmont fault system. This kinematic change appears to be reflected in the clastic wedges by a Morrowan (Early Pennsylvanian) unconformity. Partitioning of deformation between shortening in the Valley and Ridge and strike-slip motion in the eastern portion of the orogen is likely related to dextral transpression between Laurentia and Gondwana. In the eastern portion of the orogen, both thrusting and younger strike-slip motion were accompanied by medium-grade metamorphism and plutonism that are spatially related to the major Alleghanian fault zones.

In most of the northern Appalachians, Alleghanian deformation is manifested mainly by dextral strike-slip faults. Sedimentation was generally localized in narrow elongate basins associated with these faults, and multiple unconformities in the basins attest to synkinematic deposition. The Alleghanian Orogeny is attributed to the oblique collision of Laurentia with Gondwana, associated with the assembly of Pangea, the Late Paleozoic supercontinent. Clearly, in the northern Appalachians the event was more of a "grazing" of Laurentia by Gondwana, whereas in the southern Appalachians, the two crustal blocks collided more head-on, although the partitioning of strain in the southern Appalachians attests to the transpressive nature of the collision there.

Just as the Grenville Orogen served as one of the seams along which Rodinia broke up, the Appalachian Orogen formed the locus of Mesozoic rifting that led to the breakup of Pangea and the formation of the modern Atlantic. Elongate basins containing rift facies clastic sedimentary rocks and mafic magmatic rocks are preserved along the length of the orogen, much as in their ancestor Iapetan rift basins.

22.4.5 Closing Remarks

The orogen has provided fodder for many tectonic concepts and continues to lure us with the many stones still unturned in its mountains, hollows, and coves. Some of the first-order observations and questions that arose as I composed this essay are:

- Does the New York promontory mark the end of a major transform in the Iapetus Ocean—perhaps one that split the ocean into two major domains, as reflected in the difference in accretionary history between the northern and southern Appalachians?
- The Taconic Orogeny, one of the oldest events in the orogen, is well preserved in the north but strongly overprinted by the Alleghanian Orogeny in south. These relations suggest that during the Middle and Late Paleozoic, accretion in the northern Appalachians mainly involved a strong strike-slip component and that areas of intense Salinian and Acadian deformation and metamorphism were localized collisions (at the scale of the orogen) where strike-slip motion was impeded by promontories.
- The traditional "mantra" of "Taconic, Acadian, Alleghanian" is giving way to the realization that tectonic activity was ongoing along the Laurentian margin and that it is somewhat naive to view the development of the orogen exclusively in the time frames of three named events.
- The nature of Late Mesozoic and Cenozoic erosional and epeirogenic events that are responsible for the modern form of the mountain range are not well known and await the attention of future researchers.

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22.5 THE CALEDONIDES—An essay by Kevin T. Pickering[1] and Alan G. Smith[2]

22.5.1 Introduction

In this essay we review the early Paleozoic history of the **European Caledonides** from Svalbard (Spitsbergen) through the northwest European and British Caledonian belts. These orogenic belts mark the edges of lithospheric plates and originated from crustal stresses associated with the subduction of former oceans, and the collision of the adjacent continents. In the Early Paleozoic, or Cambrian to Silurian (~ 570–408 Ma) the present-day Caledonides region consisted of essentially three large continental blocks separated by one or more oceans (Figure 22.5.1): Gondwana (South America and Africa being the most important continents in relation to the Caledonides), Laurentia (North America, Greenland, and northwestern Scotland), and Baltica (northwestern Europe to the Ural Mountains in the east, and south to the poorly defined southern edge in the region of the Tornquist Teisseyre lineament, which stretches from the North Sea—via the Polish Caledonides and its concealed eastward extension—to the Urals).

Three collisional belts formed between these three continents. The first, the Caledonides of Norway, western Sweden, and eastern Greenland, lies between western Baltica and eastern Laurentia; the second, a poorly exposed branch of the Caledonides under the North Sea, continues into eastern Europe situated between northern Gondwana and southern Baltica; the third consists of the Caledonides of the British Isles between northwestern Gondwana and southern Baltica. Each collisional belt represents the site of a former ocean. The name "Iapetus Ocean" has been given to the ocean area in general. For convenience, the ocean between Baltica and Laurentia will be referred to as the Eastern Iapetus, that between Gondwana and Baltica as the Tornquist Ocean, and the ocean between Gondwana and Laurentia as the Western Iapetus.

In detail, the histories of the collisions are complex. In particular, continental slivers rifted away from the northern margin of Gondwana, moved away from it across the Iapetus Ocean, and collided with the margins of Laurentia and Baltica before Gondwana itself collided with these continents. In Europe, the best defined of these slivers are Eastern Avalonia (southern Britain and much of France); in North America, they are western Avalonia (the Avalon Peninsula and Gander Zone of Newfoundland, New Brunswick, and Nova Scotia), and the Piedmont Terrane and Carolina Slate Belt in the southern Appalachians. Thus there was a stage in which short-lived new oceans lay between the Gondwanan fragments and Gondwana itself. These new oceans can be regarded as parts of the Iapetus in the broad sense, but one of them, the Rheic Ocean, is highlighted in this account. It lay between the northern margin of Gondwana and parts of central and western Europe to the south of eastern Avalonia, and is discussed in the following account.

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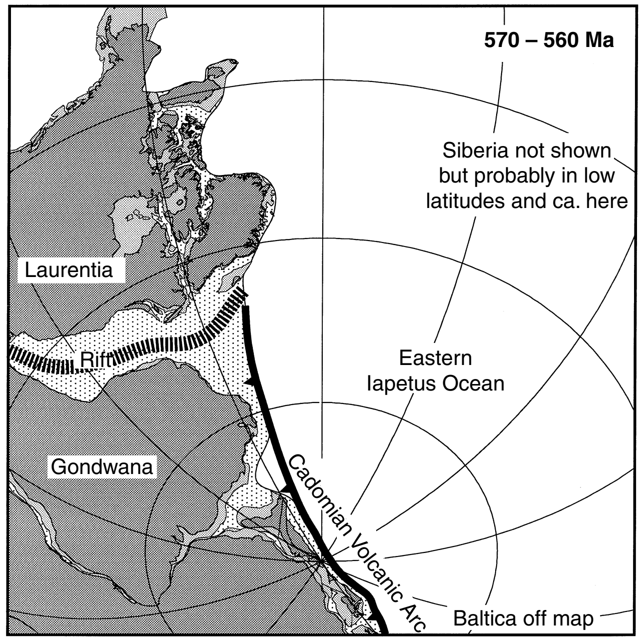
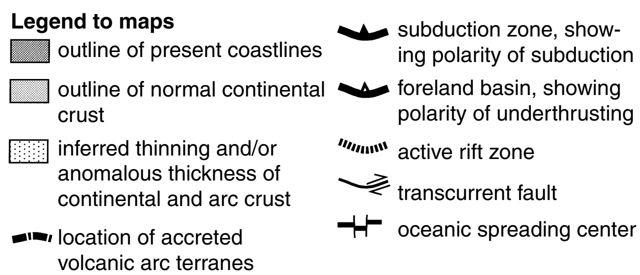
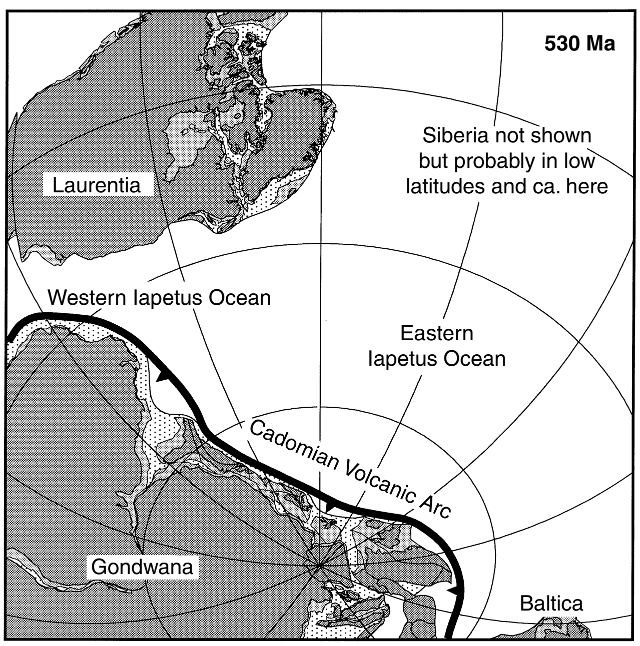
FIGURE 22.5.1 Computer-generated plate reconstructions, based on a synthesis of paleomagnetic data, for the following time intervals: Late Precambrian (570–560 Ma); Cambrian (530 Ma); Early Ordovician (490–480 Ma); Middle Ordovician (460 Ma); latest Ordovician (440 Ma); Middle Silurian (420 Ma), and early Middle Devonian (390 Ma). Positions of major magmatic arcs are shown schematically. The initial Late Precambrian reassembly joins western South America to eastern Laurentia (present coordinates). Baltica's position is uncertain and is not shown on this map. The positions have been obtained by interpolating between this initial reassembly and positions suggested by Ordovician paleomagnetic data. The paleomagnetic evidence suggests that during Cambrian time Gondwana approached Laurentia, and that northwestern South America possibly collided with an oceanic arc or arcs that fringed southeastern Laurentia in Early Ordovician time. Gondwana then rotated anticlockwise in later Ordovician to Early Devonian time, bringing opposite one another those parts of Gondwana and Laurentia (e.g., Florida) that were to collide in the later Paleozoic. Uncertainties in Baltica's position relative to other continents in Cambrian to Early Ordovician time mean that it has been omitted from maps of 490 Ma and older periods. It is first shown on the 460-Ma map (d) separated from Laurentia by a relatively narrow branch of the Eastern Iapetus Ocean. For compatibility with the events in Svalbard, the distance between the two continents is shown as decreasing between 460 and 440 Ma.



(a)





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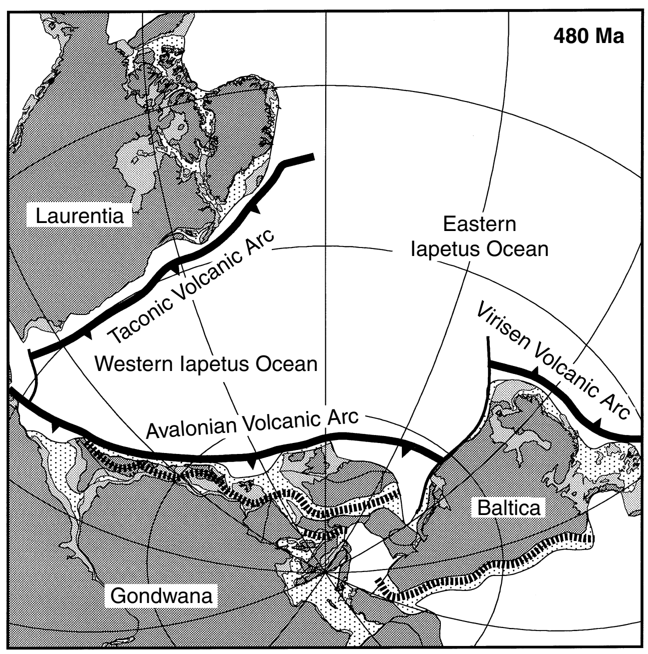

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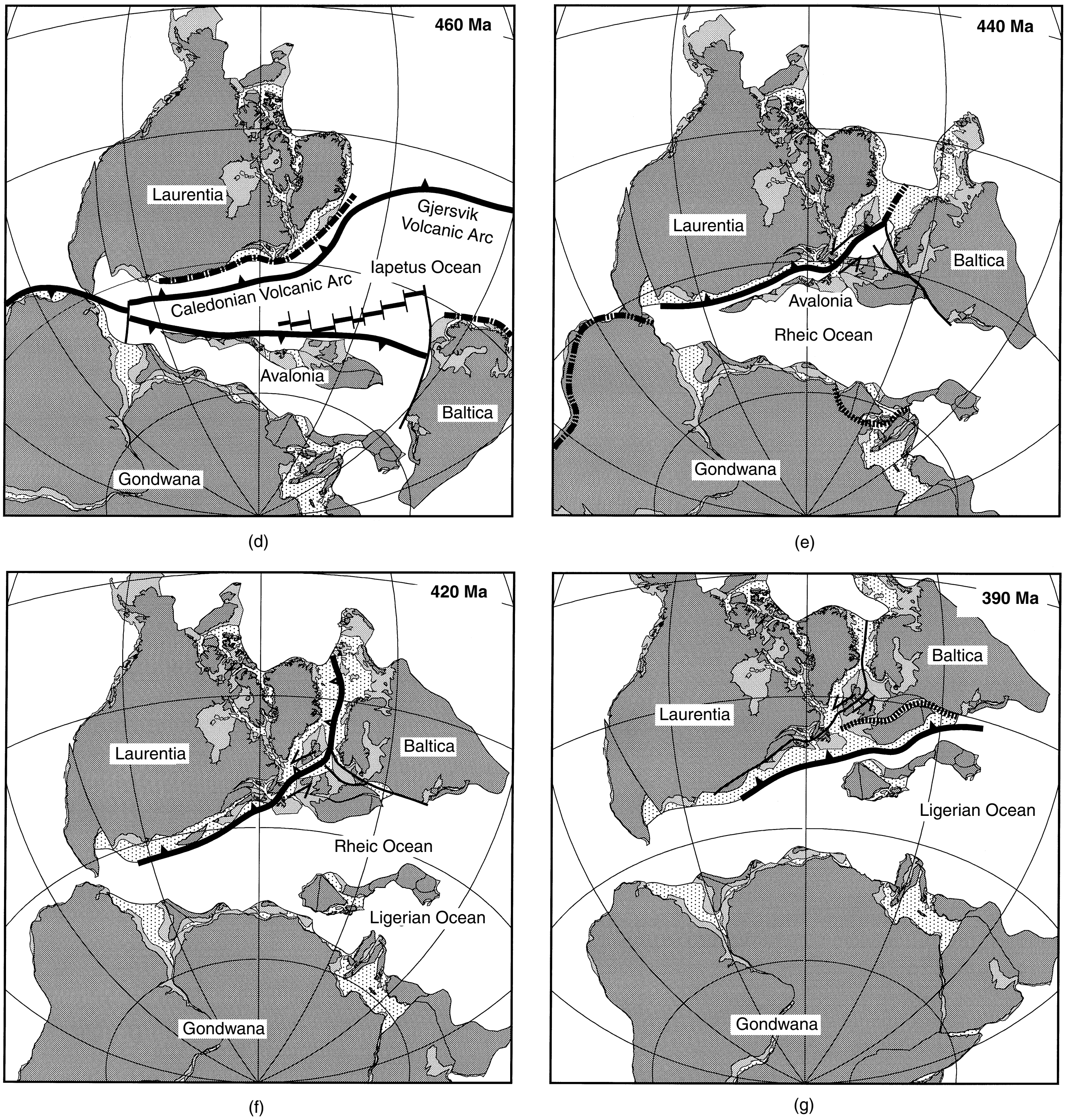

FIGURE 22.5.1 (Continued)

The orogenic belts formed by these collision events have long-established names related to their present-day geographic positions. Going back in time, the youngest phase of orogenesis (Early Devonian–Silurian) is known as Acadian in North America, Late Caledonian in East Greenland and the British Isles, and Ligerian in mainland Europe, except in Scandinavia where the term Scandian is used. In North America, the orogenic activity that peaked in the Middle Ordovician is known as Taconic, whereas in northwestern Europe, including the British Isles, East Greenland, and Scandinavia, it is referred to as the Early Caledonian; however, in Svalbard (Spitsbergen), it is known as the M'Clintock Orogeny. Late Precambrian orogenic deformation is known as the Famantinian in South America, and was for several decades known as the Grampian in northwestern Scotland (now recognized as Early Ordovician), and Cadomian in Brittany (northwestern France) and southern Britain. Although the interpretation is still controversial, we

TABLE 22.5.1 MAJOR TECTONOTHERMAL EVENTS, IAPETUS OCEAN

620–570 Ma Breakup Gondwana Phase I

Andean margin of South America rifts from eastern margin of Laurentia to create the *Western Iapetus Ocean.*

Eastern Iapetus Ocean already in existence in the Late Precambrian, with landward-dipping, peri-Gondwanan, mainly Andean-like subduction zone and subduction-related tectonothermal events, as the:

(i) *Cadomian* along northern margin of Gondwana.

(ii) *Penobscotian,* 560–510 Ma in Exploits-Gander zones of Newfoundland.

(iii) *Famantinian* in South American Andes.

(iv) *Finnmarkian* in Scandinavia, 540–490 Ma.

(v) *Grampian* in northwestern Britain, 475–460 Ma (synchronous with ophiolite obduction).

490–460 Ma Ophiolite Obduction along Northern and/or Northwestern Margin of Laurentia (Precursor to Arc-Continent Collision)

Probably due to major change from transtensional to transpressional motion between Laurentia and South America, following short-lived generation of oceanic crust in peri-Laurentian and peri-Gondwanan marginal basins.

490–470 Ma Breakup Gondwana Phase II

Avalonia (including Carolina Slate Belt, Piedmont, parts of Nova Scotia and New Brunswick, Avalon Peninsula, southern Britain) and other microcontinental terranes rift away from Gondwana. Late Arenig rift event along entire Urals.

Peak ~450-Ma Orogenesis Involving Arc Collision(s)

Taconic in U.S. and Canadian Appalachians, 480–440 Ma (arc-continent collision along eastern margin of Laurentia).

Early Caledonian in northern Britain, 530–430 Ma (arc-continent collision along eastern margin of Laurentia).

From ~470 Ma, collision of magmatic arc(s) and western margin of Baltica as it rotates around to collide with Laurentia during Silurian time.

M'Clintock in Svalbard, 500–450 Ma (arc-continent collision).

435–370-Ma Destruction of Iapetus Ocean and Associated Events

Late Caledonian, 460–380 Ma (Baltica-intervening arc(s)-Laurentia collision).

Scandian, 430–400 Ma (Baltica-Laurentia collision).

Ligerian, 390–370 Ma (collision of Gondwana-derived Aquitaine-Cantabrian blocks with eastern Avalonia-Baltica).

suspect that the Late Precambrian events represent lithospheric extension rather than orogenesis. We have endeavored to show what is meant by the terms in Table 22.5.1, which summarizes the time span for the events, their location, and cause. The end result is that remnants of small continental fragments of island arcs and of backarc basins that originally bordered the continental margins have been swept up and incorporated into all these orogenic belts as "terranes" and ophiolites during the closure of these branches of the Iapetus Ocean.

The processes that led to the eventual welding of the three supercontinents to form Pangea in Permian time gave rise to a perfect match: there are no gaps in the reassembly, such as unfilled remnant ocean basins. The continents are most unlikely to have matched perfectly before collision and one must therefore expect processes to have taken place that allow imperfectly matching continents to fit together, such as thickening of continental crust, indentation by promontories, and lateral movement of slices along "escape structures." One might also anticipate that the area around the triple junction in the North Sea, where all three continents meet, is likely to be one of the more tectonically complex areas and one of long-lived activity.

Paleomagnetic poles from the stable parts of the three large continents provide the principal quantitative data for repositioning the continents relative to one another in early Paleozoic time. The data are poor, particularly for Cambrian time (see text to follow).

Fortunately, the Late Precambrian to Early Cambrian was a time when new continental margins formed along Laurentia, Baltica, and parts of Gondwana. By joining what are believed to be the original opposing margins together it is possible to make a plausible reassembly of the circum-Iapetus continents, and this reassembly serves as a starting configuration for its evolution through the remainder of early Paleozoic time. Where paleomagnetic evidence is inconclusive, geologic evidence permits an independent inference about the nature of each continental margin from its stratigraphic record. The geologic evolution of the area is therefore discussed next and is related to the maps where possible.

22.5.2 Late Precambrian–Cambrian Extension and Passive Margins

The Late Precambrian–Cambrian history of Laurentia, Baltica, and western South America appears to reflect Late Precambrian breakup and the development of passive continental margins. Continental breakup of the Late Precambrian supercontinent of Gondwana was diachronous. In northwestern Britain, the mainly Late Precambrian Dalradian Supergroup accumulated between Laurentia and South America (Gondwana). In West Africa, the earliest recorded tectonic event is the westward rifting of a continental fragment from the West African craton about 700 Ma, and the development of a rift-drift stratigraphy (including tholeiitic and alkaline basalts). Dyke swarms, reflecting crustal extension, are known along the Laurentian margin in Labrador dated at 615 Ma. Similar dyke swarms are known from the Baltica margin in Scandinavia, dated at 665 Ma and ~640 Ma. In themselves, the dykes merely indicate stretching, but in all the areas studied they pass upward into, or are closely associated with, the development of extensive carbonate platforms (believed to have accumulated in low latitudes), and/or were intruded into very thick deep-water clastics, that are interpreted as passive continental margin successions. The eastern Laurentian passive margin extended from Greenland through the Durness sequence of northwestern Scotland to fringe most of the United States. Similarly, the Late Precambrian–Cambrian margin of Baltica is interpreted as a 200-km wide passive margin sequence. A comparable passive margin of the same age is known in northwestern Argentina and the western margin of the South American craton.

Margins formed by lithospheric extension have a characteristic subsidence curve determined both by the amount of stretching and the time it began. From such curves, the breakup and rifting ages of the new passive margins of Laurentia, Baltica, and northwestern Argentina occurred between 625 and 555 Ma, in agreement with all the other evidence. Also, in western Newfoundland, a rift-drift transition has been proposed for western Newfoundland at about 570–550 Ma.

Eastern North America (Laurentia) and western South America (Gondwana) separated from one another in the Late Precambrian to form opposing margins. This suggestion is supported by the otherwise puzzling distribution in northwest Argentina of early Cambrian *olenellid* trilobites that are similar to those in eastern Laurentia and not known elsewhere. Rift events recorded in dyke swarms from Baltica and Greenland suggest that rifting occurred about 100 million years earlier than in the Appalachians, so that the Eastern Iapetus Ocean was much older than the Western Iapetus Ocean.

22.5.3 Late Precambrian–Cambrian Arcs, Northern and Northwestern Gondwana

By contrast with the passive margins of Gondwana, the Late Precambrian–Cambrian history of northern and northwestern Gondwana is one of arc formation and orogenesis. The orogeny, known as the Cadomian, is a Late Precambrian–Cambrian (650–500 Ma) belt exposed on the northern and northwestern edge of Gondwana, with its type area in Brittany, northwestern France. It includes arc and arc-related rocks that record 650–500 Ma age tectonothermal events, also exposed in southern Britain, Spain, southeastern Ireland, and the "Avalonian" of the northern Appalachians. To the east of Brittany, Cadomian deformation is recognized in Czechoslovakia and the south Carpathian-Balkan region, so this was a time of extensive new crustal growth. The Cadomian events probably represent the vestiges of a major arc system formed by subduction under the northern and northwestern margin of Gondwana.

Many areas affected by the Cadomian Orogeny were later detached from the margin of Gondwana and migrated across the intervening ocean to become "exotic" terranes attached to Laurentia, before Gondwana itself collided with them and Laurentia. Thus, during the Late Precambrian, what were to become exotic microcontinental blocks, such as eastern and western Avalonia, the Piedmont Terrane, and the Carolina Slate Belt, probably formed the outboard parts of Gondwana along a margin that bordered an existing ocean.

22.5.4 Early to Middle Ordovician Arcs, Marginal Basins, and Ophiolites

Throughout northwestern Europe, regional chemical and isotopic signatures in the Ordovician-Devonian igneous suites north of the Iapetus suture, or its inferred along-strike continuation, show a subduction-related affinity associated with both southeast- and northwest-directed subduction, which was active from Early Ordovician to Middle Silurian times. By late Tremadoc to early Arenig time (~490 Ma) the Laurentian craton was fringed by marginal basins and arcs. Two subparallel arcs appear to have existed along much of the western/northwestern margin of the Iapetus Ocean (Figures 22.5.2a and b): (1) an inboard island-arc system that was developed mainly on continental crust, (2) a second island-arc system occupying a more oceanic setting, and developed above a north- to northwest-dipping subduction zone associated with the destruction of the Iapetus Ocean *sensu stricto*. This latter arc system appears to have been slightly younger in age and locally lasted into Silurian time. Destruction of the marginal basins took place throughout the Early Ordovician to Late Silurian interval along most of its length, but was restricted to Middle Ordovician time in Svalbard (M'Clintock Orogeny) and to latest Ordovician–earliest Silurian time in central Newfoundland (Taconic Orogeny).

A Taconic (~470–460 Ma) high-pressure granulite event has been identified in the Moine Supergroup in

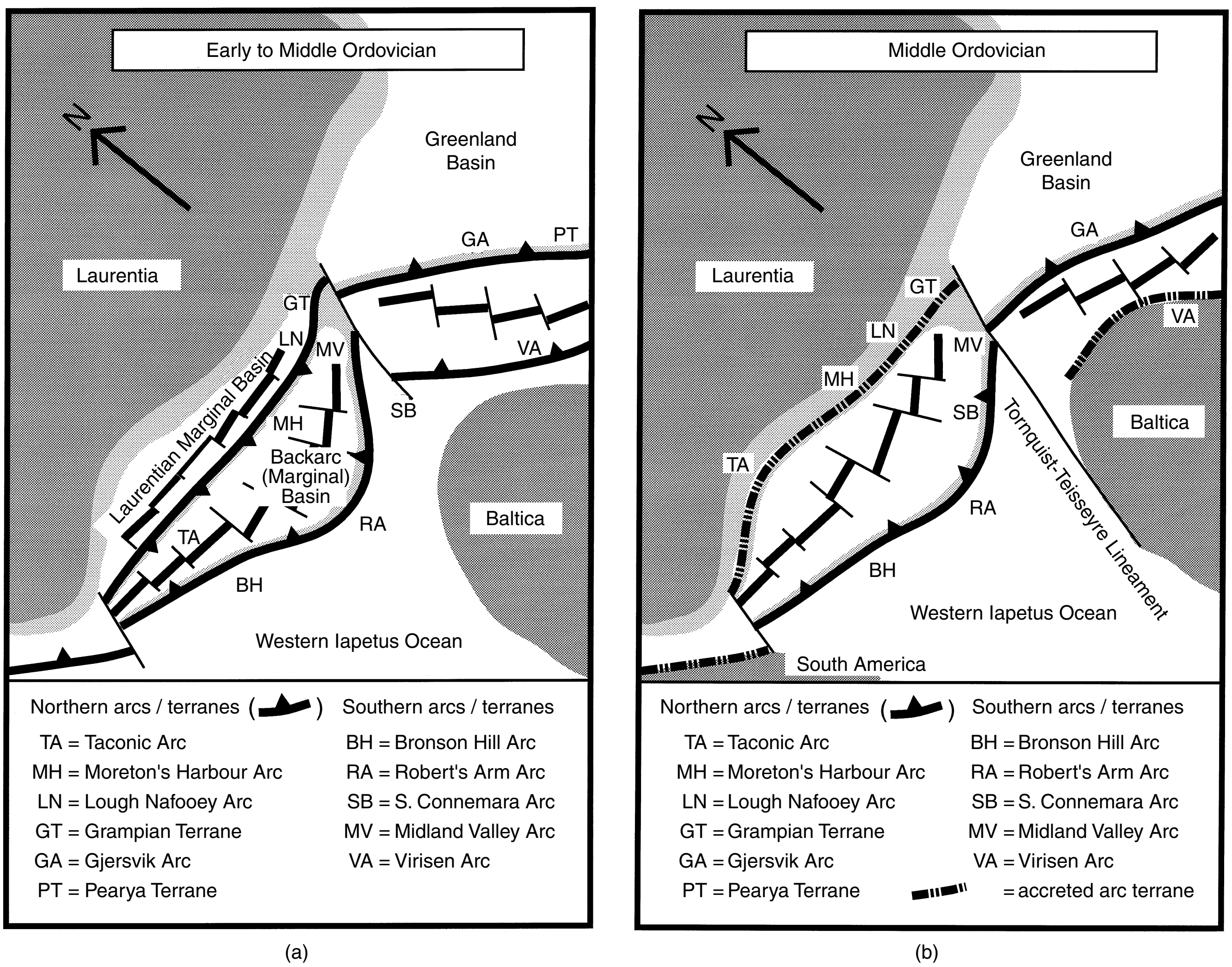

FIGURE 22.5.2 Schematic diagram showing Early to mid-Ordovician arcs and marginal basins in the Iapetus Ocean. These were subsequently accreted, in part, to Laurentia and South America (Taconic-Grampian-Early Caledonian Orogeny) and to Baltica (final emplacement in Scandian [i.e., Late Caledonian] orogeny). Note that the main Scandinavian Caledonide arcs are shown as originally located more oceanward of the Taconic arc(s). Solid black arrows indicate the probable polarity of subduction; light gray areas indicate thinned continental crust. Not to scale.

northern Scotland. U-Pb geochronologic data from Connemara, in the western Irish Caledonides, suggest that continental arc magmatism along the southern Laurentian margin was short-lived, lasting from ~475 to ~ 463 Ma. A corollary of these ages is that the Grampian Orogeny in Connemara was considerably younger than is generally acknowledged. The Grampian Orogeny, therefore, was synchronous with the Taconic Orogeny in the northern Appalachians.

Like the Mesozoic Tethys, the Western Iapetus Ocean was associated with a relatively brief phase of arc development and oceanic spreading, rapidly followed by plate convergence leading to extensive, Early Ordovician ophiolite obduction along the Laurentian margin immediately preceding arc-continent collision (Taconic-Grampian–Early Caledonian). From west to east, early Ordovician ophiolites fall into three categories: (1) Laurentian marginal basin oceanic crust, as in western Newfoundland and the Shetland Islands; (2) intra-arc, backarc marginal basin oceanic crust, as in the central Newfoundland, western Ireland, and Highland Border Group ophiolites (and possibly in slivers along the Highland Boundary fault zone), and (3) forearc ophiolites, including accreted seamount material (as in eastern central Newfoundland and southwest Scotland), which may include seamount fragments. In the Scandinavian Caledonides, arc fragments appear to have been involved in collisional events in Early Ordovician times (Finnmarkian), but were finally emplaced onto Baltica during the Silurian–Early Devonian Scandian orogeny.

The marginal basins that fringed Laurentia in the Early Ordovician were closed by the end of Middle Ordovician time. In North America their closure caused extensive east-west shortening (today's coordinates), mainly accommodated by westward transport of thrust slices (Taconic Orogeny). The Taconic Orogeny of eastern North America lasted from about 480 to 430 Ma, with a peak between 470 and 450 Ma. The Taconic Orogeny overlaps in time with the Early Caledonian (M'Clintock) events in Svalbard (Spitsbergen) on the west side of Baltica.

The closure of the marginal basins that fringed Laurentia in the Early Ordovician was caused by arc-continent collision, driven by the proximal approach of the northern margin of Gondwana (probably northern South America) to the Appalachian margin of Laurentia, and/or a major reorientation of relative plate vectors because of plate-tectonic processes associated with other ocean basins. The collision events along the Laurentian margin (including the Grampian Orogeny) on the northern side of the Eastern Iapetus Ocean, may be related to the narrowing of the ocean between Baltica and Laurentia such that, by the late Llandeilo, generic faunal links were established across a Galapagos-like island chain between both continents.

The maps for this period (Figure 22.5.2, 480–460 Ma) show western South America moving past eastern Laurentia. The motion appears to have been quite oblique, without actual collision, but caused arc accretion in Middle Ordovician time in the absence of an intense Himalayan-style continent-continent collision between Laurentia and western Gondwana. South America (or "Occidentalia"), which is bordered by a Cambrian carbonate platform similar to that of eastern North America, may have been a continental margin opposing Laurentia during Late Ordovician time. In northwest Argentina, the early Paleozoic *olenellid* trilobite faunas in the Famantinian Orogen are similar to those found in eastern Laurentia, which may suggest geographic linkage between these areas.

As noted previously, the position of relative to other continents, is uncertain, therefore it has been omitted from maps of 490 Ma and older periods. It is first shown on the 460-Ma map, where it appears separated from Laurentia by a relatively narrow branch of the Eastern Iapetus Ocean (Figure 22.5.2). For compatibility with the events in Svalbard, the distance between the two continents is shown as decreasing between 460 and 440 Ma. This presumed narrowing of the ocean between Baltica and Laurentia is supported by the establishment of faunal links between the two continents by late Llandeilo time (about 465 Ma), where none had existed previously.

Throughout northwestern Scotland and northern Ireland, north of the Iapetus suture, geologic data, including regional chemical and isotopic signatures in the Ordovician-Devonian igneous suites, are associated with northwest-directed subduction that was active from Early Ordovician to Middle Silurian times. This subduction zone appears to have been active in accommodating the subduction of most of the Iapetus oceanic crust, although southeast-directed subduction may have been important locally, and even in the final stages of closure.

22.5.5 Early Ordovician Breakup of the Northwest Margin of Gondwana

In the late Arenig, there was a second major episode of continental fragmentation in Gondwana (Phase II breakup, to distinguish it from the Late Precambrian Phase I breakup. The Avalonian, Piedmont, and Carolina Slate Belt Terranes, all of which contain Cadomian-age arc basement, broke away from the northwestern edge of the Gondwana continent. The partial separation of eastern Avalonia (including northwestern France, also called Armorica) from Gondwana during the late Arenig, has been documented using sedimentological criteria, including subsidence curves, from the Sahara, Middle East, Nova Scotia, Ibero-Armorica, Ireland, England, and Wales.

In addition to the separation of these fragments, a new ocean basin may have been created on the eastern edge of Baltica in the present region of the western Urals. A more speculative view is that the Eastern Iapetus Ocean (i.e., the Tornquist Sea) was created at about the same time by the separation of southern Baltica from northern South America, that is, Phase II breakup could have resulted from Baltica rifting away from Gondwana along with Avalonia, and the Piedmont and Carolina Slate Terranes. After rifting, Baltica may have moved toward the equator, from a more southerly position in Early Ordovician time. There is no clear evidence for the creation of new passive margins in these areas like that for those of Late Precambrian to Cambrian age discussed previously. Avalonia is not linked with Baltica at this time.

22.5.6 Middle–Late Ordovician Subduction, Continental Fragmentation, and Collisions

The late Arenig Phase II breakup of Gondwana was probably temporally linked with the extensive ophiolite obduction initiated in the Llanvirn. Much of the northern Gondwanan margin of the Iapetus Ocean was an Andean-type covergent plate margin with a subduction polarity towards the continental interior. Remnants of this arc include the calc-alkaline igneous rocks of the English Lake District, southern Welsh Basin, and southern Ireland. The Welsh Basin was initiated as a marginal basin on the southern side of the Eastern Iapetus Ocean during the Arenig, at the same time as eastern Avalonia rifted off northwestern Gondwana. Because the late Arenig Phase II breakup of Gondwana immediately preceded the extensive ophiolite obduction initiated in the Llanvirn, both processes may be causally related.

During Cambrian to Early Ordovician time, northwestern Britain was part of Laurentia and located at about 15°–20°S. By contrast, the paleolatitude of eastern Avalonia (southern Britain) in Early Ordovician time was about 60°S. The paleolatitude of eastern Avalonia had changed to about 45°S in the Middle Ordovician, and ~15°–25°S in the latest Ordovician–Early Silurian, suggesting it had a steady northward drift across the Iapetus Ocean. The latitudinal separation across the Iapetus Ocean in the Early Ordovician, between the part of Laurentia containing northwestern Scotland (where eastern Avalonia eventually docked), and the Gondwanan margin with eastern Avalonia, changed from about 5000 km in the late Tremadoc–early Arenig to ~3300 km by the Llanvirn-Llandeilo. The underlying plate-tectonic causes for this northward motion are not immediately obvious. The paleomagnetic data independently support the faunal arguments for the northward movement of eastern Avalonia across the Iapetus Ocean during the Late Ordovician.

Subduction-related igneous activity occurred in the British Isles, south of the Iapetus suture, from the Tremadoc (earliest Ordovician) to earliest Caradoc (Middle-Ordovician) with a rapid change to a more alkaline and peralkaline signature and abrupt cessation in the Longvillian (Middle-Caradoc). Ridge subduction, and the creation of a slab window below the northern margin of Gondwana, (i.e., Eastern Avalonia; see Figure 22.5.1d), provide an elegant mechanism to explain (a) the abrupt switch-off in subduction-related igneous activity in eastern Avalonia in the early Caradoc; (b) the changed geochemical signature of the Caradoc compared with earlier igneous activity in eastern Avalonia; (c) the subduction of thermally warm ridge-flanks millions of years prior to ridge subduction as a reason for the widespread Llandeilo hiatus, or thin stratigraphies, throughout much of eastern Avalonia, and (d) a fundamental cause for the transference of eastern Avalonia to a north-moving plate and eastern Avalonia's rifting away from Gondwana, as the ridge is spreading center jumped southward of the microcontinent. An analogy can be found in the present-day Pacific where the small continental fragment of Baja California is now attached to the Pacific Plate and moving with it.

The protracted collision events contemporaneous with the Taconic Orogeny of North America culminated in high-grade metamorphism and major uplift at about 460–440 Ma in the Western Iapetus Ocean, and along other parts of the Laurentian margin and the associated marginal basins. Parts of the Scandinavian Caledonides (Figure 22.5.3) record an uplift in 450–435 Ma. Stable argon isotope studies of rocks from northern Sweden in the Upper Allochthon (Lower Koli Nappe, the Seve-Koli shear zone, the Seve Nappe) and the shear zones of the Middle Allochthon, reveal high-grade metamorphism and associated deformation of the Seve units as a Late Cambrian–Early Ordovician event in which the rocks cooled below the respective closure temperatures for hornblende at ~490 Ma and muscovite at ~455 Ma. The structurally

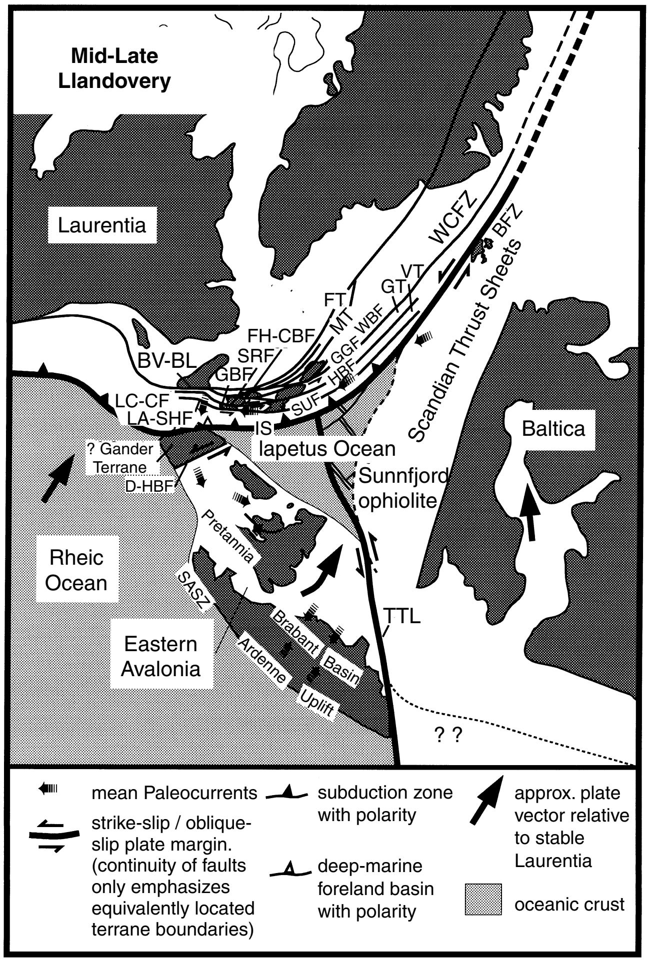

FIGURE 22.5.3 Mid-late Llandovery reconstruction. Plate motion arrows are shown for the Baltica and eastern Avalonia Plates relative to a fixed Laurentian (North American) plate. Closely spaced stipple outlines parts of Armorica, Britain, and the Gander Terrane south of the Iapetus suture, and fragments of western Newfoundland and northwestern Britain with Laurentian crustal affinities prior to ocean closure. Abbreviations for major geologic features: BV-BL = Bay Verte-Brompton Lineament, LC-CF = Lobster Cove-Chanceport Fault, LA-SHF = Lukes Arm-Sops Head Fault, GBF = Galway Bay Fault, SRF = Skerd Rocks Fault, FH-CBF = Fair Head-Clew Bay Fault, SUF = Southern Uplands Fault, HBF = Highland Boundary Fault, GGF = Great Glen Fault, WBF = Walls boundary Fault, FT = Flannan Thrust, MT = Moine Thrust, IS = Iapetus suture (i.e., Cape Ray-Reach Fault in Newfoundland); BFZ = Billefjorden fault zone (Svalbard), WCFZ = Western Central Fault Zone (Svalbard); TTL = Tornquist Teisseyre Lineament; SASZ = South Armorican Shear Zone, D-HBF = Dover-Hermitage Bay Fault.

Paired Ordovician arc systems in the northerly parts of the Iapetus Ocean correlated from Newfoundland to Svalbard and separated by ophiolites are as follows: (1) Taconic island arc, Early–Middle Ordovician developed above oceanward-dipping subduction zone, Moreton's Harbour arc (Newfoundland), Lough Nafooey arc (Ireland), Grampian Terrane (Scotland), and Gjersvik arc (Norway), Pearya/Northwestern Svalbard (including Biskayerhalvoya) Terranes; (2) Southern island arc, Ordovician-Silurian arc developed above Laurentia-ward dipping subduction zone, Bronson Hill arc, Robert's Arm arc (Newfoundland), South Connemara arc, Midland Valley terrane (Scotland) Virisen arc (Norway), and western Svalbard.

lower rocks of the Middle Allochthon, inferred to have been more proximal to Baltica prior to emplacement, show only the ~430-Ma event(s), whereas the Upper Allochthon records the older Finnmarkian event(s). In the Seve Nappes, there is evidence for Middle–Late Ordovician 450–440-Ma shear zones, showing that a pre-Scandian deformation affected rocks outboard from, or marginal to, Baltica. During this time, Seve Nappes of different P-T-t (pressure, temperature, time) histories were juxtaposed. Subsequently, during the Scandian Orogeny, the Seve and Koli Nappes were juxtaposed, and the Middle Allochthon mylonites formed as these nappes were emplaced over the Baltic Shield. All these tectonic units were assembled prior to regional cooling through the Ar closure temperature of muscovite.

22.5.7 Middle Ordovician–Silurian Closure of the Eastern Iapetus Ocean

The Ordovician-Silurian history of the Midland Valley, Scotland, records the evolution of an arc and backarc basin. Throughout the Late Ordovician and Early Silurian, the Southern Uplands of Scotland and the along-strike Wexford-County Down area (Longford Down inlier) of Ireland were part of an active accretionary prism developed above a northward-dipping subduction zone on the northern margin of the Iapetus Ocean. For example, the Southern Uplands accretionary prism, which developed over at least 50 m.y. from the Llanvirn to Wenlock, was associated with the Midland Valley forearc basin further to the north. The arc massif of older metamorphic basement in the Grampian Highlands northwest of the forearc basin was capped by calc-alkaline arc volcanics and intrusive igneous suites, and supplied most of the sediments to the trench-forearc accretionary system to the south.

Closure of the Eastern Iapetus Ocean by oblique (overall sinistral) collision took place between the island arc(s) sandwiched between the converging continents of Laurentia and Baltica. The closure may have begun, during the latest Llandeilo to Caradoc in the region of northern Norway-Svalbard, to incorporate M'Clintock-Finnmarkian orogenic crustal fragments. In places the setting probably resembled that between mainland Southeast Asia and northern Australia today.

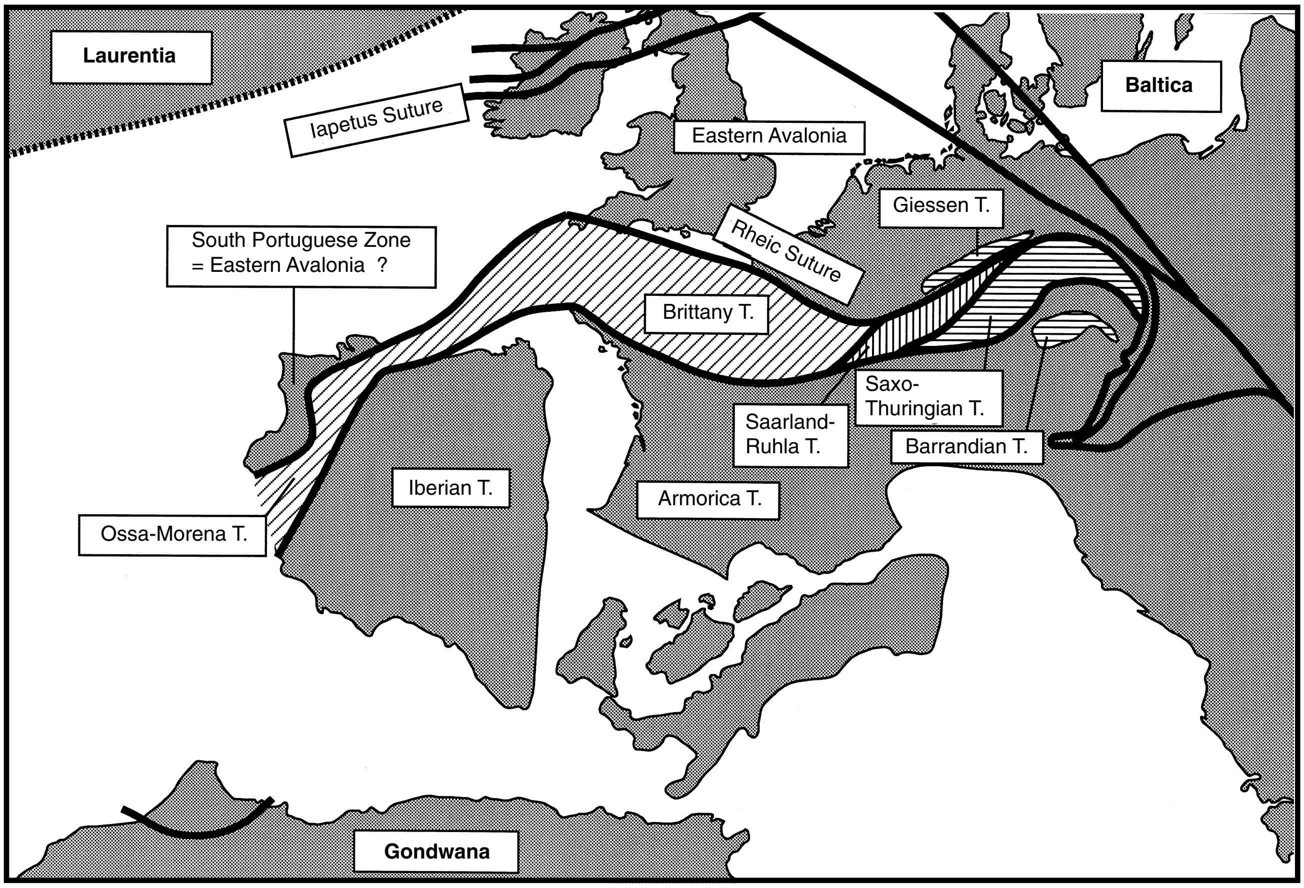

FIGURE 22.5.4 Principal European continental blocks during the Lower Paleozoic, and their suture sites (heavy lines).

Elsewhere, continent-continent collision may have created a situation like that between the present-day Himalayas and the Bengal Fan of eastern India. Voluminous flysch sediments were shed away from the collision zone to form the axial-trench wedges of sandy turbidites preserved in the Ordovician tracts of the Southern Uplands accretionary prism in northwestern Britain, akin to the present-day Bengal Fan.

The Silurian-Devonian Scandian Orogeny, caused by the collision of Laurentia and Baltica above a northwest-dipping subduction zone, resulted in the final emplacement of thrust sheets eastwards onto the Scandinavian crystalline basement with its Cambrian-Ordovician shelf successions.

Collision of eastern Avalonia with Baltica and Laurentia occurred in the latest Ashgill–earliest Llandovery, with the microcontinent behaving as a rotating rigid indentor, probably in the region of present-day central Newfoundland, above a north-dipping subduction zone (Figure 22.5.4). Collision was oblique (or "soft") with a sinistral component along the margin. The major phase of bimodal Silurian magmatism in central Newfoundland, New Brunswick, south-central Britain, and western Ireland implies that there was a component of extension shortly after the initial collision. Extension is also suggested by the kinematic history of syn-deformation granites in northwestern Britain. Even faunal evidence from Middle Silurian ostracodes suggests a phase of extension after which oblique convergence continued. The phase of Middle Silurian oblique extension was probably caused by the rotation of eastern Avalonia against Laurentia during the final stages of suturing. The main sinistral displacement of eastern Avalonia took place throughout the Silurian and up until the peak Acadian deformation in the Emsian (Early Devonian time).

Final welding of eastern Avalonia took place in the Wenlock and was associated with a prolonged phase of deep-marine foreland-basin development. By Llandovery time, the Tornquist-Teisseyre lineament, formed a major, probably active, backarc strike-slip fault to the arc associated with closure of the Eastern Iapetus Ocean between Baltica and eastern Avalonia as a submarine (subshelf) lineament. The southern continental plate boundary of Baltica remains poorly defined, but was south of the arc complex associated with northward subduction in the northwestern European Caledonides.

Silurian-Devonian sinistral shear was associated with the amalgamation of the Western, Central, and Eastern Provinces of Svalbard. These provinces have pre-Devonian histories that involved complete separation of the terranes. Eastern Spitsbergen and Nordaustlandet, for example, may have originated along the Laurentian margin far to the south of their present position, to be juxtaposed against the Central Province along the Billesfjorden fault zone by the Late Devonian. In Svalbard, argon ages suggest an Early Devonian cleavage formation at ~400 Ma.

In north-central Newfoundland, the age of the slaty cleavage is constrained as latest Silurian (Ludlow-Pridoli), whereas in eastern Avalonia the slaty cleavage appears to range in age from Early Silurian to early Middle Devonian. In Wales and the Welsh Borderland, the Middle Devonian is commonly missing, with postorogenic sedimentation resuming in the Late Devonian (Famennian). Also, in north Devon (southwestern England), the early Middle Devonian involved the influx of clastics derived from the Welsh Basin, which is interpreted as substantial uplift (orogeny) to the north. In the Midland Valley, Scotland, the Middle Devonian is also missing, but it was the main period of chiefly fluvio-lacustrine sedimentation in the Orcadian Basin, northeastern Scotland.

Within the Caledonian slate belt, south of the Iapetus suture, there is a major arcuate trend in the orientation of the strike of cleavage, from an Appalachian (northeast) trend in Ireland and Wales to a more Tornquist (east-southeast) trend that is typical of northern Germany and Poland, with intermediate trends in northern England. Based on the clockwise cleavage transection of related fold axial surfaces, and associated sinistral displacement on strike-slip faults in northwestern England, there was an episode of Late Caledonian ("Acadian") sinistral transpression. The microgranite dyke swarm associated with the emplacement of the Shap granite intrudes folded and cleaved Silurian sediments, but the dykes themselves are weakly cleaved. In the English Lake District the formation of this cleavage was contemporaneous with the emplacement of various syn- and post-deformational igneous suites, dated at about 394–392 Ma or early Middle Devonian (Emsian). The cleavage arcuation (from north-northeast in northwestern England to a more easterly trend further east in northern England) is explained as a consequence of the anticlockwise rotation of eastern Avalonia relative to Laurentia during collision and final suturing.

22.5.8 Late Ordovician Icehouse

In Late Ordovician time (Hirnantian stage of the Ashgill ~443–444 Ma), a major short-lived glaciation affected western Gondwana (mostly northwest Africa). The icehouse may have lasted for as little as 0.5 m.y., or it may have lasted 4–6 m.y., beginning in the Caradoc, with peak glaciation during the Hirnantian stage of the Ashgill. Stable-isotope data from brachiopods show a dramatic positive isotope excursion in the $\delta^{13}C$ and $\delta^{18}O$ record (PDB scale ~2‰) for eastern North America, central Sweden, and the Baltic States. In the Baltic States, the magnitude of these isotopic excursions is up to ~4‰, which is equivalent to the combined effects of a sea-level fall of 100 m and a drop of 10°C in tropical sea surface temperatures.

Why this icehouse came into being in the middle of a greenhouse period and then lasted for such a short time has always been a puzzle. It has been proposed that its initiation and demise are attributable to the action of a Central American or similar oceanic gateway. This moved some continents into polar latitudes; while there, they opened and closed oceanic low-latitude gateways that changed global oceanic circulation from one with important circumequatorial currents (greenhouse) to one with inhibited circumequatorial deep-water currents (icehouse) and more restricted oceanic gyres. The latter scenario, favored the export of moisture to high polar latitudes, where it could accumulate as snow and ice to form the continental ice sheets on the polar parts of Gondwana.

The paleomagnetic data suggest that northwestern Gondwana and southwestern Laurentia were closer to one another at 440 Ma than they were just before or just after the icehouse. Thus there was a potential gateway in the correct period; it could have closed in the Late Ordovician and could have re-opened shortly afterwards. Plate-tectonic motions have the appropriate scales of time and length to account for the movements required: about 100 km/m.y. Also, zircon and monazite U-Pb data, together with tectonic mapping and petrologic studies in the Acatlan Complex, southern Mexico, have led workers to invoke a Late Ordovician–Early Silurian continental collision orogeny, which they ascribe to the collision of eastern Laurentia with western South America along the entire Andean margin. Thus, there appears to have been more than one important site at which oceanic gateways were active at the time of the Late Ordovician icehouse Earth.

22.5.9 Ordovician-Silurian Magmatic Arcs Elsewhere in Europe

In contrast to the Lower Paleozoic rocks of Norway and Sweden, rocks of this age elsewhere in continental Europe are largely concealed beneath younger strata or have been overprinted by intense Variscan (Hercynian or Armorican) and Alpine deformation. Exposure is poor and tectonic boundaries are difficult to define. Nevertheless, the Ordovician-Silurian outcrops of the European Caledonides can be assigned to three continental plates (Figure 22.5.4): (1) eastern Avalonia (considered above); (2) the southern parts of Baltica; and (3) parts of the northern margin of Gondwana (including, from west to east, the Brittany, Saarland-Ruhr, Tepla-Barrandian, Saxo-Thuringian, Bavarian, Gory Sowie, and Moravo-Silesian Terranes, and, farther south, the Ibero-Armorica-Moldanubian Terranes).

By Late Ordovician time, the Eastern Iapetus between eastern Avalonia and Baltica (also known as the Tornquist Sea) had probably closed by eastward and/or northeastward subduction. The southern margin of eastern Avalonia is marked by a major Early Paleozoic suture zone with ophiolites in Ibero-Armorica and Alpine Europe. The zone also includes arc-related igneous-volcanic suites and thick marine successions that can be traced through Iberia, into Armorica (along the South Armorican Shear Zone, or SASZ), and across to southern Austria. The SASZ includes thrust-bound slices of metasedimentary and igneous rocks metamorphosed to eclogite facies, 420–375-Ma blueschists, and a high-temperature low-pressure migmatite belt, interpreted as remnants of an accretionary complex formed above a subduction zone, which was almost certainly active during the Silurian. There was also a major Late Ordovician–Silurian (450–415 Ma) high-pressure event with little deformation, in France, Iberia and Morocco.

In the Ossa Morena zone, central Iberia, the Middle Devonian emplacement of northeast-vergent nappes was followed by major sinistral transpression. This created the central Iberian fold belt separating the Aquitaine-Cantabrian microcontinent to the east from the South Portuguese block to the west; the latter can be regarded as belonging to the southern part of eastern Avalonia.

Further east, from Early Silurian to Middle Devonian time, the mid-European and Tepla-Barremian "terranes" appear to have been subject to considerable crustal extension, and the extrusion of voluminous, within-plate, alkali basalts. In the eastern Alps of southern Austria, there are subduction-related volcanics and sediments that formed in an island arc and active continental margin setting. These are probably vestiges of an arc on the edge of Gondwana, but their precise relation to Gondwana is unclear.

The Ligerian Orogeny involved the collision of these Gondwana-derived fragments with the southern margin of eastern Avalonia, which was itself a fragment broken off Gondwana at an earlier period. It involved the development and destruction of a volcanic arc complex, with a backarc marginal basin to the north, by continent-continent collision in the younger Variscan (Hercynian) Orogeny. A Middle Devonian intermediate-pressure metamorphic event associated with major tectonism is well known, not only in France, but also in Morocco and, more speculatively, in Iberia. This event is interpreted as reflecting the amalgamation of other continental fragments with eastern Avalonia, such as Saxo-Thuringia and Moldanubia, in addition to those in Aquitaine and Cantabria.

The Rheic Ocean lay to the south of the Ligerian orogenic belt and north of Gondwana, that is, south of the Tepla-Barremian Plate, where the Ibero-Armorica-Moldanubian Plate was converging northwards throughout the Devonian. Essentially, the Rheic Ocean is synonymous with the vestiges of the Eastern Iapetus Ocean. The ocean appears to have been closed by Middle Devonian (Givetian) time, though its closure may have been preceded by the creation of a small backarc basin to the north. All oceanic areas between Gondwana and Laurussia (Laurentia and Baltica) had been eliminated by Late Carboniferous (Namurian) time. Subduction of the Rheic Ocean crust probably was initiated in the Early Devonian, and ocean closure occurred mainly by the northward subduction of oceanic crust below the southern arc-related margin of the Baltica Plate (which includes cratonic Russia west of the Urals, a Permian collisional orogen).

The southeastern margins of some of the microcontinental terranes that were accreted to Laurentia-Baltica during the Ordovician-Devonian are marked by Late Devonian ophiolites; for example, the obduction of the ~400–375-Ma Lizard ophiolite around 370 Ma. After unfolding the major Variscan flexure from Armorica to Iberia, and allowing for significant dextral offset along faults, such as the fault that displaces the Haig Fras from the rest of the Cornubian Batholith, it appears that the approximately 390-Ma Morais ophiolite, one of a series of crystalline complexes exposed in northern Iberia, represents the along-strike equivalent of the Lizard complex. These ophiolites were obducted northwards as fragments of the Rheic or Ligurian Ocean crust.

22.5.10 Postorogenic Continental Sedimentation and Igneous Activity

The collision of Baltica with Laurentia united them into a single continent known as Laurussia and created a high mountain chain. The collision of Gondwana and Laurussia to form Pangea was not completed until Late Carboniferous time. Molasse accumulated in fault-controlled intermontane basins, which were preserved along the major tectonic lineaments and suture zones.

By the Early to Middle Devonian, most of the major strike-slip between terranes appears to have occurred along many lineaments, for example, the Great Glen Fault of northwestern Scotland is a major strike-slip fault but does not appear to have been significantly active during Old Red Sandstone (ORS) deposition. Furthermore, the Emsian-Eifelian Orcadian ORS next to the fault shows a net offset today of 25–29 km, which has a dextral sense, rather than the sinistral sense implied by the geometry of docking and collision. Post-ORS dextral offsets are also known in the Shetland Isles and are much larger—on the order of 120 km.

The high topography was supported by thickened continental crust. Temperatures in the lower part of the crust rose sufficiently to partially melt it and produce late-stage granites and granitoid bodies. These are the late-orogenic and postorogenic intrusions that characterize the final stages of continent-continent collisions. By earliest Carboniferous time, the crust on the southern margin of Laurentia was extending and spreading laterally by gravitational collapse, which induced faulting at the surface. Such faults should be at right angles to the greatest stress, that is, parallel to the mountain chain.

The tectonic vergence divide between the major northward (Late Paleozoic coordinates) obduction and thrusting events associated with closure of the Western Iapetus Ocean, and the southward tectonic transport onto the Baltic Shield due to closure of the Eastern Iapetus Ocean, was situated in the region of central Newfoundland to the British Caledonides. This latter region probably was the site of a triple junction associated with transforms and spreading centers during the opening of the Western and Eastern Iapetus Oceans, evolving into a triple junction involving subduction zones and transforms. The obliquity of the collision events may have been a major contributory factor in the preservation of the low-grade slate belts, in contrast, a Himalayan-style collision, which occurred farther north, resulted in high-grade metamorphic rocks being common at the surface.

22.5.11 Closing Remarks

The Late Precambrian–Early Cambrian was marked by the opening of an essentially east-trending Western Iapetus Ocean between southeastern Laurentia and western South America. The Eastern Iapetus Ocean, separating the Greenland area of Laurentia from Baltica opened earlier than the western ocean and was elongated in a north-south direction. Both oceans had different histories of opening and closure. Ocean closure involved overall sinistral shear between Baltica and Laurentia during the Late Ordovician to Devonian, whereas the U.S. Appalachians were influenced by dextral shear (first transpressive, then transtensional) between South America and Laurentia during the Ordovician.

Like the Mesozoic breakup of Pangea, the breakup of Gondwana (Phase I, ~620–570 Ma; Phase II, ~490–470-Ma) may have been associated with plume activity though, the evidence for this is not clear. Continental breakup would probably have increased the length of the global ocean-ridge system and caused a reduction in the mean age of the ocean floor. This would have resulted in a global rise in sea level, leading to a widespread flooding of continents. This is consistent with the preponderance of wide Cambro-Ordovician shelf seas, which were commonly sites for the accumulation of organic-rich muds (now pyrite-rich black shales). The lack of glaciogenic sediments or striated pavements, at least until the latest Ordovician (Ashgill), suggests that there were no substantial, if any, polar ice caps. The Cambrian-Ordovician was probably a greenhouse period induced by enhanced atmospheric CO_2 levels, which, in turn, are attributable to increased oceanic-ridge and mantle-plume activity.

Future research needs to better constrain the timing and nature of basin-forming events, and their subsequent histories and destruction, a task that can be accomplished through improved radiometric dating techniques and careful (macro- to micro-) structural and stratigraphic and/or sedimentologic studies. Careful reevaluation and new measurements of paleomagnetic data, for example, those using stepwise thermal demagnetization techniques, are giving us an improved understanding of the movement history of continental fragments. Sophisticated geochemical and isotopic arguments are helping to define the plate-tectonic settings of igneous rocks (supra-subduction zone, extensional intraplate, etc.), and to infer past global and/or regional climates from sediments. Undoubtedly, the Lower Paleozoic remains an area of fruitful research.

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22.6 TECTONIC GENEALOGY OF NORTH AMERICA—An essay by Paul F. Hoffman[1]

22.6.1 Introduction

Continents are complex tectonic aggregates that evolved over hundreds of millions of years. Today's continents are drifted fragments of the mid-Phanerozoic **supercontinent** Pangea, more or less reshaped by accretion and ablation at subduction zones. The creation of Pangea involved the fusion of many older continents, each of which was itself a drifted fragment of some still older continental assembly. Orogenic belts, marking the sites of ocean opening and closing, are the basic elements that are used to unravel the tectonic genealogy of continents.

Today's giant continent, Eurasia, was assembled in the Phanerozoic eon (545–0 Ma) through the piecemeal convergence of many pre-Phanerozoic continental fragments, roped together by Phanerozoic subduction complexes. The assembly of Eurasia is ongoing today. Other continents are fragments of former giant continents assembled at various times in the pre-Phanerozoic. The southern continents, for example, are derived from Gondwanaland, which was assembled in the Neoproterozoic era (1000–545 Ma). North America (Figure 22.6.1) is the largest fragment of a continent assembled in the Paleoproterozoic (2500–1600 Ma); other fragments exist in Eurasia and probably elsewhere. Since the Paleoproterozoic, North America (Laurentia, exclusive of fragments lost to Europe when the Atlantic opened) has twice collided to form supercontinents (all continents gathered together). The older collision is represented by the Mesoproterozoic (1600–1000 Ma) Grenville Orogen and the resulting supercontinent, named Rodinia, had an approximate age span of 1050–750 Ma. The younger supercontinent is Wegener's Pangea, which had an age span of 300–150 Ma. It was conjoined with North America along the Appalachian orogen and its connecting orogenic belts around the Gulf of Mexico (Ouachitas), East Greenland (Caledonides), and Arctic Canada (Franklin). Ancestral North America, therefore, participated in most of the salient tectonic events of the past three billion years.

The Phanerozoic evolution of the continents is reasonably well understood and key Phanerozoic orogens are the subjects of other essays in this book. Earth's pre-Phanerozoic history is less well known, despite having produced over 80% of existing continental crust, but it is a subject in healthy ferment. Before discussing the role of ancestral North America in pre-Phanerozoic continental evolution, I should clarify some terminology. A collisional orogen implies a fusion of mature (>200 million years old) continental blocks. An accretionary orogen implies the addition of juvenile (<200 million years old) oceanic material—accretionary prisms, magmatic arcs, and volcanic plateaus, for the most part. For brevity, I will use the numerical geon time scale, where geon 0 equals 0–99 Ma, geon 1 equals 100–199 Ma, geon 10 equals 1000–1099 Ma, and so on. The divisions of the Proterozoic eon, defined above, are those recognized by the International Union of Geological Sciences and differ slightly from the parallel divisions in the Geological Society of America DNAG time scale.

[1]Harvard University, Cambridge, MA, USA.

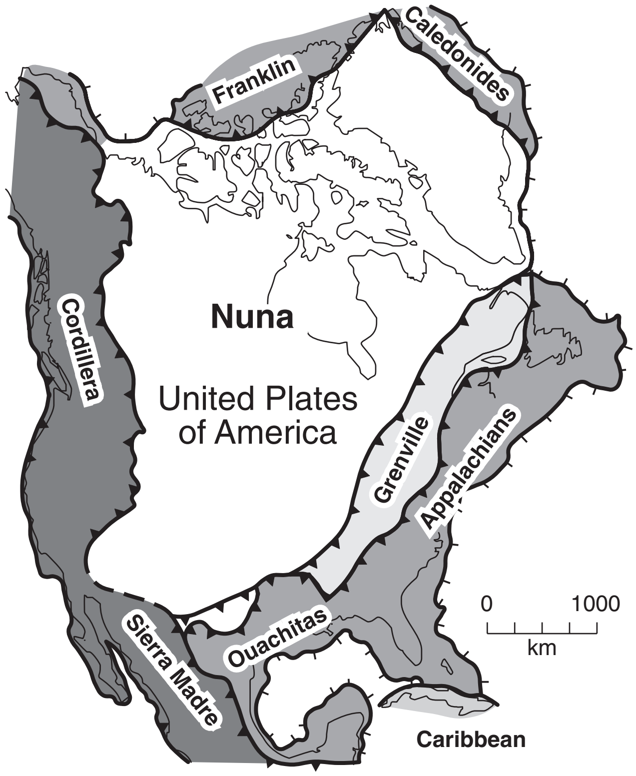

FIGURE 22.6.1 Simplified orogenic structure of North America: a Paleoproterozoic nucleus, Nuna, is discontinuously bordered by the Meosproterozoic Grenville and Racklan (far northwest) Orogens; the Paleozoic Ouachita, Appalachian, Caldonide, and Franklin Orogens, and the Mesozoic-Cenozoic Cordilleran and Caribbean Orogens. Greenland is restored to pre-rift (>90 Ma) position.

22.6.2 Phanerozoic (545–0 Ma) Orogens and Pangea

The stable interior of North America is framed by two great Phanerozoic orogenic systems. The Cordilleran system (Figure 22.6.1) borders the Pacific Ocean basin and is still active. The Pacific continental margin first opened in geon 7, but the main phase of tectonic accretion occurred in geon 1, coeval with rapid northwesterly drift of the continent (relative to hot spots) following the breakup of Pangea and opening of the North Atlantic Basin. Cordilleran crust that was thickened during Mesozoic accretion collapsed in extension in the Paleogene, when convergence between North America and the Pacific Basin slowed. Dextral strike-slip deformation became increasingly important in the Neogene, when North America began to override the East Pacific spreading ridge. The Cordillera will continue to be active until the Pacific Basin closes. If the Atlantic continues to open, North and South America will eventually collide with eastern Asia, which by then will have incorporated Australasia.[2]

The other Phanerozoic orogenic system formed during the Paleozoic assembly of Pangea. It includes the Appalachian, Ouachita, Caledonide, and Franklin orogenic belts (Figure 22.6.1). They evolved from a continuous continental margin that opened diachronously in geons 6 and 5. Parts of the margin collided with island arcs and became active margins in geon 4 and, by the end of geon 3, the northern Appalachian and Caledonide segments had collided with Baltica, and the southern Appalachian and Ouachita sectors had done the same with northwest Gondwanaland. The resulting orogenic system was dismembered when the North Atlantic Basin opened.

22.6.3 Neoproterozoic (1000–545 Ma) Orogens and Gondwanaland

Gondwanaland (Figure 22.6.2), the former giant continent that broke up in the Mesozoic to form Africa, South America, Antarctica, Australasia, and southern Eurasia, was assembled in the Neoproterozoic era. Gondwanaland was an aggregate containing at least five older continents—West Africa, Amazonia, Congo, Kalahari, and East Gondwanaland (Australia, East Antarctica, India). They were welded together by a network of Neoproterozoic collisional orogens and bordered by Neoproterozoic-Paleozoic accretionary orogens. Pre-Phanerozoic North America lacks Neoproterozoic orogens, but displaced continental slivers that originated on the northwest margin of Gondwanaland were incorporated into the Appalachians in Middle and Late Paleozoic time. A two-way land trade apparently occurred in the Middle Paleozoic, suggesting a glancing encounter between eastern North America and western South America. Part of the southern Appalachians ended up in northwest Argentina and a strip of northern South America (the Avalon Terrane) was added to the coast of New England and eastern Canada. Later, a piece of northwest Africa (the Florida Peninsula and panhandle) was transferred to North America during the climactic Late Paleozoic collision with Gondwanaland.

[2]The new supercontinent is dubbed Amasia.

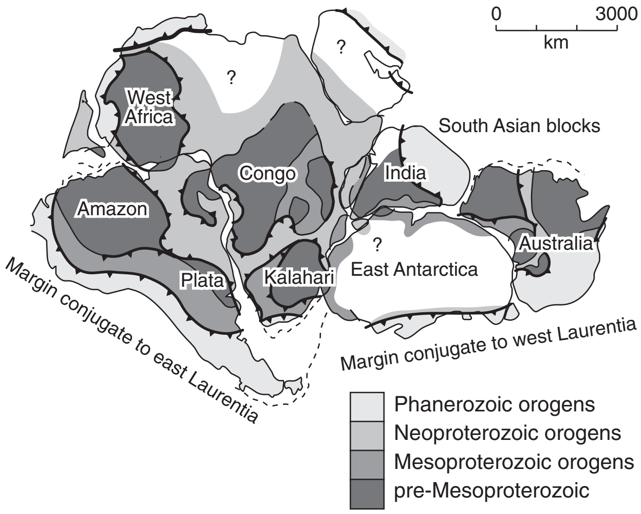

FIGURE 22.6.2 Aggregate structure of Gondwanaland, cemented by Neoproterozoic collisional and accretionary orogens. The reconstruction is well constrained by dated Mesozoic-Cenozoic seafloor magnetic anomalies. Dotted lines show edges of extensive modern continental shelves. Margins conjugate to Laurentia based on the Rodinia restoration in Figure 22.6.3. Note the disaggregated nature of Mesoproterozoic orogenic segments and the Neoproterozoic orogenic belt entering East Antarctica opposite Sri Lanka.

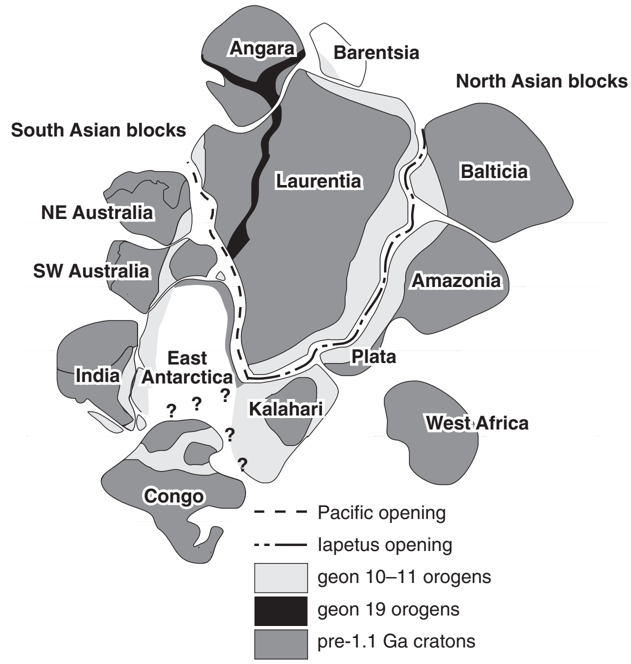

FIGURE 22.6.3 Hypothetical reconstruction of Rodinia, showing lines of Neoproterozoic opening of the Pacific and Iapetus Ocean Basins. As shown, Rodinia was an aggregate of cratons cemented by geon 10–11 (Grenvillian) orogens. Present-day north arrows for Laurentia (North America plus Rockall Bank and northwest Britain) and Kalahari cratons point to top and bottom of page, respectively.

22.6.4 Mesoproterozoic (1600–1000 Ma) Orogens and Rodinia

The Mesoproterozoic Grenville Orogen lies inboard of the Appalachians and extends for 5000 km from Mexico to Labrador (Figure 22.6.1). It truncates Archean and Paleoproterozoic structural fabrics and tectonic boundaries to the northwest, implying that the orogen evolved as a rifted or sheared continental margin. The orogen comprises an outer (northwestern) zone consisting of reactivated Archean and Paleoproterozoic basement, and an inner (southeastern) zone of juvenile Mesoproterozoic crust. The outer zone is characterized by northwest-directed crustal-scale thrust shears and exposes metamorphic rocks that underwent >30 km of post-1050-Ma exhumation. Thrusting occurred in geons 11–10 and is presumably related to accretion of the inner zone and terminal collision of ancestral North America with an outboard continent(s). The hypothetical Grenvillian hinterland must have broken away when the Iapetus paleocean basin opened, initiating the Paleozoic Appalachian orogenic cycle. Likely candidates for the Grenvillian hinterland are the Amazonia-Plata craton of South America (Figure 22.6.2) and the Baltica craton of northern Europe. Both are flanked by Grenville-age orogenic and plutonic belts, consistent with those cratons belonging to the overriding plate of the terminal Grenvillian collision.

Rifted segments of orogenic belts of Grenville age occur throughout Gondwanaland, except for the West African craton (Figure 22.6.2). In addition to the belts in South America mentioned previously, deeply eroded geon 11–10 orogenic belts occur in central and southern Africa, including Madagascar; in Sri Lanka and the Eastern Ghats of India; and in eastern Antarctica and Australia (Figure 22.6.2). Under current investigation is the hypothesis that all or most of these segments originally belonged to a continuous, 10,000-km-long system analogous to the Neogene Alpine-Himalayan system. The product of these collisions was the supercontinent **Rodinia** (Figure 22.6.3), the exact configuration of which is still conjectural, but which should make a structurally compatable restoration of the Grenvillian orogenic segments. The Rodinia reconstruction (Figure 22.6.3) implies that Gondwanaland was turned inside-out following the breakup of Rodinia at about 750 Ma. The requisite anticlockwise rotation of East Gondwanaland, relative to Laurentia, and convergence with West Gondwanaland before about 500 Ma is consistent with paleomagnetic data.

22.6.5 Paleoproterozoic (2500–1600 Ma) Collisional Orogens and Nuna

The vast region between the Grenville and Cordilleran Orogens (Figure 22.6.1), including the marginal parts of the orogens, was assembled in Paleoproterozoic time. It incorporates at least four Archean microcontinents—Churchill, Superior, Nain, and Slave (Figure 22.6.4). Paleomagnetic data indicate >4000 km of Late Paleoproterozoic convergence between the Churchill and Superior cratons, broadly contemporaneous with significant relative motion between the Churchill and Slave cratons. The Wyoming craton may be an extension of the Churchill or a fifth independent microcontinent; the mutual boundary is buried by thick Phanerozoic platform cover. The Churchill has two major divisions—Rae and Heane (Figure 22.6.4)—previously thought to have fused in the Paleoproterozoic. However, recent studies indicate an earlier, Late Archean time of assembly. The overall Paleoproterozoic assembly has been called the United Plates of America, and is commonly believed to be continuous with Baltica (Figure 22.6.4). An appropriate name for the entire continent assembled by the end of the Paleoproterozoic is **Nuna,** an eskimo name for the lands bordering the northern oceans and seas.

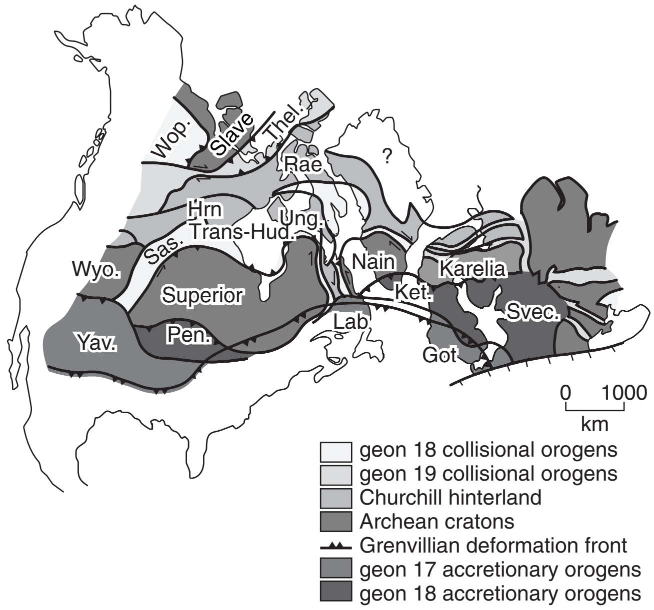

FIGURE 22.6.4 Existing aggregate structure of Nuna: the United Plates of America and its extension in Baltica after Gorbatschev and Bogdanova (1993). Nuna was cemented by geon-18 orogens, which are truncated at the present margins of Nuna, and by the peripheral geon-17 accretionary orogens. Other extensions of Nuna exist on other continents or have been destroyed. Got = Gothian Orogen; Hrn = Hearne craton; Ket = Ketilidian orogen; Lab = Labrador Orogen; Pen = Penokean Orogen; Sas = Saskatoba syntaxis; Svec = Svecofennian Orogen; Thel = Thelon Orogen; Trans-Hud = Trans-Hudson Orogen; Ung = Ungava syntaxis; Wop = Wopmay Orogen; Wyo = Wyoming craton; Yav = Yavapai Orogen.

The subduction zones that accommodated the converging Archean microcontinents dipped predominantly beneath the Churchill continent, with far-reaching structural and magmatic consequences. The Churchill margins have well-developed Paleoproterozoic plutonic belts, representing the eroded roots of continental magmatic arcs (Figure 22.6.5). These are lacking on the Churchill-facing margins of the Superior, Nain, and Slave cratons. The Churchill continent was far more severely and extensively deformed by the collisions than were the cratons. Large-scale strike-slip and oblique-slip shear zone systems developed as weak Churchill crust was extruded laterally in response to indentation by the three more-rigid cratons (Figure 22.6.4). The Churchill continent also experienced unique intraplate magmatic events during and after the collisions. Ultrapotassic alkaline volcanism occurred in

FIGURE 22.6.5 The charnokitic Cumberland batholith in central Baffin Island is a product of Paleoproterozoic arc magmatism on the margin of the Churchill hinterland. Mount Asgard, the nunatak in the center of the photo, rises over 800 meters above the surrounding glacier.

a 100,000-km^2 area west of Hudson Bay close to the time of the Churchill-Superior collision. Almost 80 million years later, after most of the collision-related deformation and metamorphic unroofing had occurred, the same region underwent high-silica rhyolite volcanism and associated rapakivi-type granite emplacement over an area of 250,000 km^2. Analogies have been drawn between the Churchill hinterland and the Neogene Tibetan Plateau, hinterland of the Himalayan collisional orogen.

The margins of the cratons facing the Churchill are characterized by large-scale thrust and nappe structures, directed away from the Churchill hinterland. Paleoproterozoic sedimentary and volcanic rocks deposited on the rifted margins of the cratons are discontinuously preserved and exposed. Volcanism and dike swarms related to initial continental breakup occurred mainly in geons 21–20. Subsequent passive-margin sediments (platformal carbonates and mature fine clastics) are overlain disconformably by foredeep sediments (in complete ascending sequence: ironstones, black shales, greywacke turbidites, and redbeds). The change occurred as the leading edge of each craton entered a peri-Churchill subduction zone. The stratigraphic transition and hence the onset of collision can be precisely dated if suitable material (e.g., air-borne volcanic ash layers) is present.

The structurally higher levels of the overthrust belts bordering the Superior craton are composed of quasi-oceanic material. At the Ungava syntaxis (Figure 22.6.4) in northern Quebec, parts of an oceanic plateau (1.92 Ga), an imbricated ophiolite (2.00 Ga), and an immature island arc (1.87–1.83 Ga) were thrust southwards across the rifted margin onto the Superior craton. The ophiolite, one of the world's oldest, includes a sheeted dike complex, ultramafic cumulates, and volcanic suites chemically and isotopically correlated with mid-ocean ridge and ocean island basalts. At the Saskatoba syntaxis in northern Saskatchewan and Manitoba (Figure 22.6.4), the craton is tectonically juxtaposed by juvenile island-arc-type volcanic and plutonic rocks (1.93–1.85 Ga) and derived metasediments. The craton at first formed the structural footwall, but was later thrust toward the hinterland over Paleoproterozoic juvenile rocks of the intervening Trans-Hudson Orogen (Figure 22.6.4). The northeast-facing lateral margin of the craton and the arcuate embayment between the Ungava and Saskatoba syntaxes (Figure 22.6.4) contain allochthonous mafic sill–sediment complexes formed in syncollisional pull-apart basins (also called rhombochasms). The thrust-fold belts on the lateral margins of all three indented cratons are flanked by crustal-scale strike-shear zones bordering the hinterland (Figure 22.6.4). Structurally, the belts have much in common with Phanerozoic collisional orogens, but lithologically they are richer in volcanic rocks, consistent with higher mantle temperatures in the Paleoproterozoic.

The timing of the Paleoproterozoic collisions is best constrained by U-Pb geochronology of (1) the passive-margin to foredeep stratigraphic transition described earlier, (2) the cessation of arc magmatism or change from arc-type to collision-type magmatism on the Churchill margins, and (3) the exhumation of metamorphic rocks having pressure-temperature-time trajectories characteristic of collisional origin. The geochronological data show that the Slave-Churchill collision occurred first, beginning about 1.97 Ga. The Nain-Churchill collision occurred about 100 m.y. later, and the Superior craton joined the assembly at about 1.84 Ga.

22.6.6 Paleoproterozoic Accretionary Orogens Add to Nuna

Around the time the Superior, Nain, and Slave cratons collided to nucleate Nuna, juvenile Paleoproterozoic crust began to be accreted onto their trailing margins (i.e., those facing away from the Churchill hinterland). The respective accretionary orogens are the Penokean of the Great Lakes region, the Ketilidian of South Greenland and adjacent Labrador (where it is called Makkovik), and the Wopmay around Great Bear Lake, Northwest Territories (Figure 22.6.4). All three orogens evolved from passive margins that collided with island arcs and were converted to Andean-type margins as a result of subduction-polarity reversal (meaning that subduction zones first dipped away from the cratons and later beneath them). Arc-continent collision in the Wopmay Orogen occurred at 1.88 Ga, almost 90 m.y. after the Slave-Churchill collision. In the Ketilidian Orogen, chronometrically less well defined, the arc-continent collision is placed at about 1.84 Ga, about 30 m.y. after the Nain-Churchill collision. Arc-continent collision in the Penokean orogen occurred about 1.85 Ga, close to the time of the Superior-Churchill collision. The confluence of cratons and accretion at their trailing margins suggests a large-scale pattern of lithospheric flow converging on the Churchill hinterland. A possible explanation is that the hinterland was situated over a vigorous downwelling region of the sublithospheric mantle self-sustained by the descent of cold oceanic slabs.

Accretion on the (present) southern and southeastern margins of Nuna was renewed in geon 17, following apparent truncation of geon 18 structures on those margins. The accretionary orogens of geon 17 are principally exposed in Labrador and the southwestern United States, but, based on studies of drill core, they also make up the buried basement across most of the southern midcontinent (Figure 22.6.4). Juvenile crust, at least 1300 km wide, was accreted in geon 17 between the Wyoming craton and the Grenville Orogen of West Texas. The accreted material proved to be a fertile source for Mesoproterozoic crustal-melt granites and rhyolites, which are extensively encountered in the midcontinent subsurface.

The widely held belief that Nuna originally included Baltica (the Baltic Shield and East European platform) is based on proposed continuity between the Archean Nain and Karelia cratons, the Ketilidian and Svecofennian Orogens of geon 18, and the Labrador and Gothian Orogens of geon 17 (Figure 22.6.4). A more tentative connection is proposed between Nuna and the Angara craton of Siberia, based on extensions of the Slave-Churchill collision zone (Thelon Orogen) across the Arctic. Even more tenuous links exist between Nuna and major geon-18 orogenic belts in northern and western Australia. In addition, there are tantalizing similarities in Late Paleoproterozoic to Early Mesoproterozoic platform cover sequences worldwide that have contributed to the notion, as yet undemonstrated, that Nuna was a giant continent predating Rodinia.

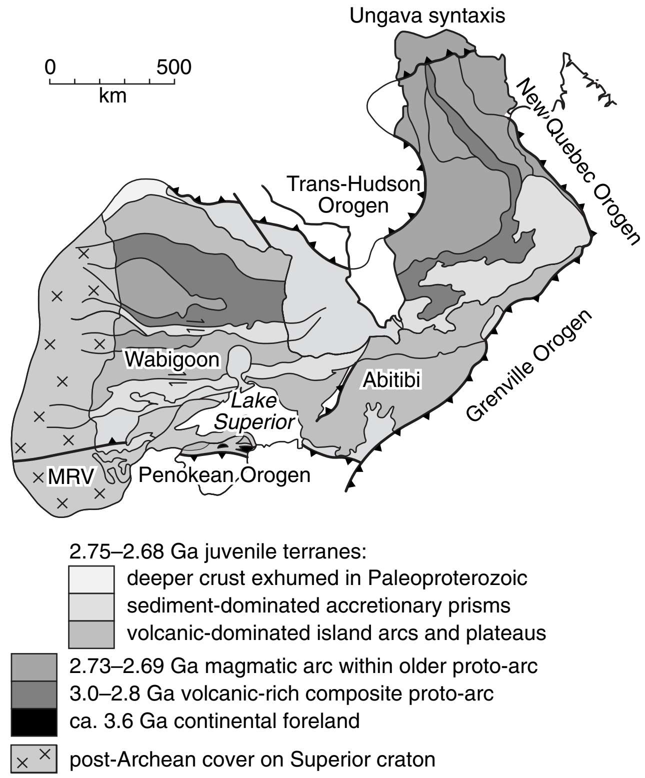

FIGURE 22.6.6 Accretionary structure of the Superior craton, the largest Archean craton in Nuna (Figure 22.6.4). Note change in tectonic strike from east-west to north-south and the truncation of tectonic boundaries at the edge of the craton. MRV = Minnesota River Valley Terrane.

22.6.7 Archean Cratons and Kenorland

Archean cratons have dimensions below which their plate tectonic settings are difficult to determine. The largest one continuously exposed is the Superior craton of the Canadian Shield (Figure 22.6.6), measuring 2500 km east to west and 1500 km north to south. It was constructed in the latter half of geon 27. It exposes, in zonally varying proportions, deformed plutonic and volcanic rocks of island-arc and oceanic-plateau affinities and derived sediments. The regional tectonic strike swings from east-west in the western and southeastern parts of the craton, to north-south in the northeastern part (Figure 22.6.6). In the northwest of the craton, a long-lived composite protoarc had evolved since geon 30. Bilateral accretion of juvenile material onto the protoarc began about 2.75 Ga. A progressively southward docking sequence of south-facing volcanic arc and accretionary prism couplets occurred until 2.69 Ga. Accretion was terminated by the collision of the entire assemblage with an old (~3.6 Ga) continent, the Minnesota River Valley (MRV) Terrane (Figure 22.6.6). The southward docking sequence and the structural evidence of persistent dextral strike-slip displacement accompanying the docking events imply overall oblique northwest-directed subduction. In the northeast of the craton, where the tectonic strike is north-south, a relatively high proportion of plutonic rocks, deep level of exhumation, and absence of strike-slip displacements indicate a less oblique subduction regime. The tectonic zonation and associated structural grain of the craton is clearly truncated at its western, southeastern, and northeastern margins (Figure 22.6.6), showing that the entire craton is a rifted fragment of a Late Archean continent, referred to as **Kenorland.**

There is a good possibility that the Wyoming craton (Figure 22.6.4) was originally connected to the south margin of the Superior craton in Kenorland. The Wyoming craton is old, like the MRV, and it preserves erosional remnants of a highly distinctive Early Paleoproterozoic sequence of uraniferous conglomerates, tropically weathered quartz arenites, and glacial diamictites that are remarkably similar to the Huronian sequence on the north shore of Lake Huron in the southern Superior craton. Separation of the Superior and Wyoming cratons occurred in the latter half of geon 21, about 300 m.y. before the two cratons were reunited in Nuna.

A remarkable number of cratons worldwide underwent major accretion or collision in geon 27. It was a time of oblique plate convergence and rapid continental accretion in the Yilgarn craton, the largest in Australia, and the time of collision between the Kaapvaal and Zimbabwe cratons of southern Africa, to name just two of the more important overseas examples. In North America, the Hearne Province of the Churchill hinterland was accreted in geon 27, but the Slave and Nain cratons were assembled in geon 26. Precise U-Pb dating of large-scale mafic dike swarms, emplaced during breakup events associated with mantle plumes, holds promise as a means of identifying formerly contiguous cratons. This should yield important insights into the extent and configuration of Kenorland, possibly the first giant continent.

22.6.8 Closing Remarks

North America consists of an aggregate nucleus that assembled in the Paleoproterozoic, discontinuously bordered by Mesoproterozoic and Phanerozoic orogens. Parts of North America participated in perhaps five giant continents: Kenorland, assembled in geons 27–26; Nuna, assembled in geons 18–17; Rodinia, assembled in geons 11–10; Gondwanaland, assembled in geons 6–5; and Pangea, assembled in geons 3–2.

The recurrence interval for giant assemblies seems to have become shorter with time—900, 700, 500, and 300 m.y. This may merely reflect our inferior knowledge of the earlier part of the record. If these intervals are real, on the other hand, how can this pattern be reconciled with secular cooling of the earth, which would result in mantle convection of diminishing vigor. It seems unlikely that continental collisions have become more frequent. However, collision zones tend to reopen, particularly when young. This tendency could have been even stronger in the distant past. To make a giant continent piecemeal, the disassembly processes must be checked. Therefore, the growing incidence of giant continents may reflect increased stability, rather than increased frequency, of continental collisions.

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22.7 PHANEROZOIC TECTONICS OF THE UNITED STATES MIDCONTINENT

22.7.1 Introduction

If you've ever driven across the United States, or have looked at a relief map of the country, you can't help but notice that the nature of topography varies radically with location. Along the East Coast, the land rises to form the long ridges of the Appalachian Mountains, whereas in the west, numerous chains of rugged peaks together compose the broad North American Cordillera. In between these two mountain belts, a region known geographically as the **Midcontinent,** the land surface is relatively flat and low-lying (Figure 22.7.1). Roads wind among dramatic cliffs and deep valleys in the mountains, but shoot like arrows across the checkerboard of farmland and rangeland in the Midcontinent. These contrasts in topography reflect contrasts in the character and geologic history of the continent's crust.

Following the definitions of Table 22.7.1, we classify the Midcontinent region as a **continental-interior platform.** Here, a cover of Phanerozoic strata was deposited in wide but shallow seaways over a basement of Precambrian (crystalline) rock. A continental-interior platform is one of two kinds of crustal provinces that can occur in a craton. We define a **craton** as continental crust that has not developed penetrative fabrics and tight folds, and has not been subjected to regional metamorphism since the beginning of the late Neoproterozoic (i.e., since about 800 Ma). In addition to continental-interior platforms, a craton can include a **shield,** a region where Precambrian basement rocks crop out extensively at the ground surface, either because cover strata were never deposited or because they were eroded away after deposition. In North America, much of the interior of Canada is a shield—not surprisingly, the region is known as the Canadian Shield.

The mountain ranges of the United States consist of two kinds of crust. Specifically, the Appalachians and the North American Cordillera can be considered Phanerozoic orogens, for during the Phanerozoic these regions developed penetrative deformation and, in places, regional metamorphism. They were also the site of widespread igneous activity and significant uplift. As a consequence of Phanerozoic tectonism, these orogens became warm and relatively weak. What is often overlooked is that the North American Cordillera also includes a large area of crust that is identical in character to that of the continental-interior platform, even though it has been uplifted significantly and has been locally subjected to deformation and igneous activity during the Phanerozoic. This crust underlies the Rocky Mountains and Colorado Plateau of the United States. In this essay, you will see that the tectonic behavior of the Midcontinent resembles that of the Rocky Mountains and Colorado Plateau, except on a more subdued scale.[1]

Cratons differ from younger orogens not just in terms of topography, but also in terms of physical characteristics such as strength, seismicity, heat flow, and crustal thickness. Specifically, orogens have less strength,[2] more seismic activity, higher heat flow, and thicker crustal roots, than do cratons. The contrast in strength reflects the contrast in heat flow, for warmer rock tends to be more plastic, and therefore weaker,

[1]Note that the Canadian Rockies, in contrast, comprise a fold-thrust belt and, as such, are part of a younger orogen.

[2]The collision between India and Asia illustrates the strength contrast between a craton and a younger orogen. India consists of old, cold Precambrian crust, while the southern margin of Asia consists of warm, relatively weak crust. During their collision, India pushed deeply into Asia.

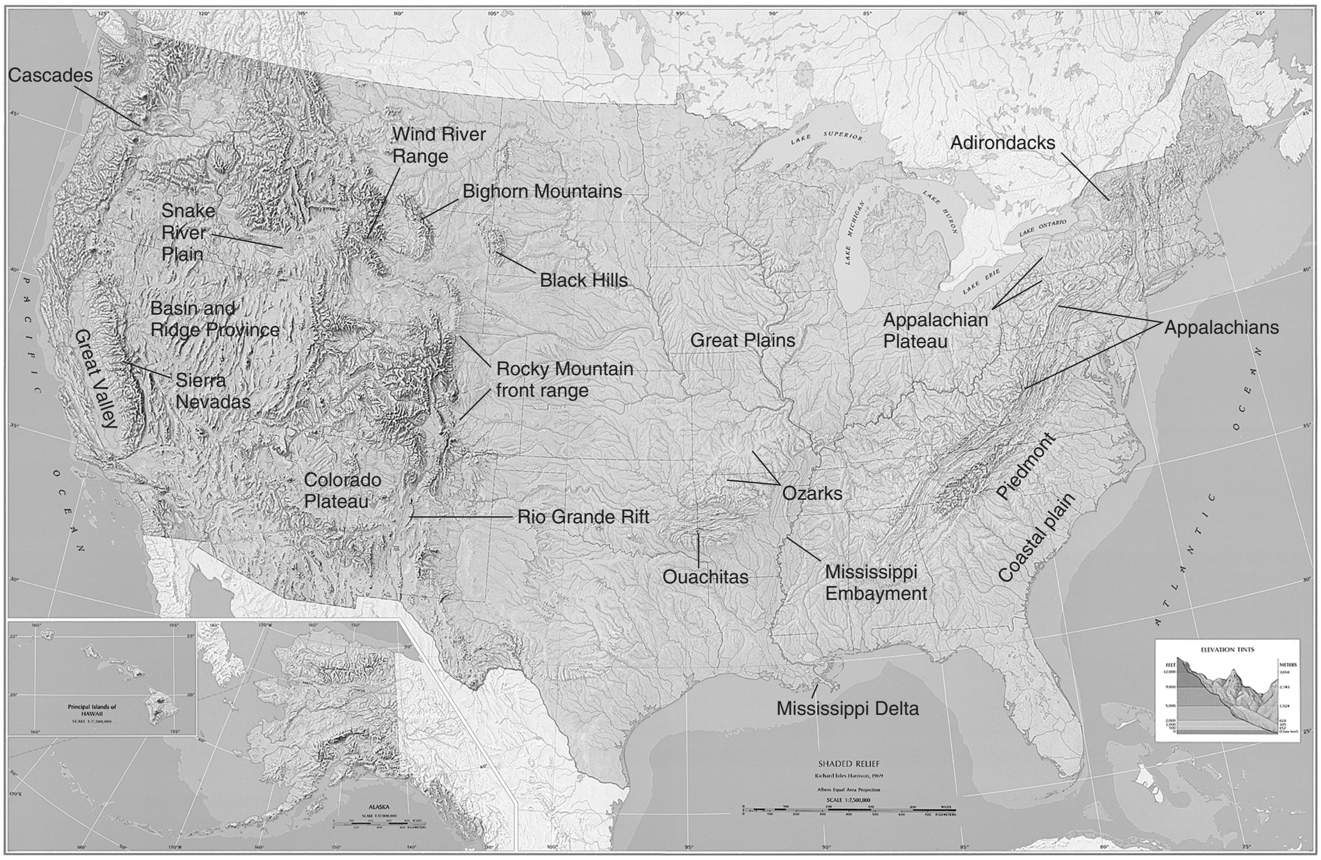

FIGURE 22.7.1 Topographic relief map of the United States, showing the contrast between Phanerozoic marginal orogens and the Great Plains.

than cooler rock of the same composition. Rock that makes up cratons initially formed either at volcanic arcs bordering convergent plate boundaries, or at hot-spot volcanoes above mantle plumes. Thus, the continental crust that eventually became the craton grew by successive collision of these buoyant blocks. This fact is well illustrated by studies in the Canadian Shield, where outcrops contain spectacular tectonite fabrics, folds, and faults that tell of a long and complex history of accretionary orogeny. If crust of the cratons began its life in orogenic belts, then how did it become strengthened and stabilized or, in other words, cratonized? Researchers suggest that cratonization comes in part from aging, because continental lithosphere loses heat and becomes stiffer as it ages. In fact, the lithospheric mantle beneath cratonic crust is thicker than that of other kinds of continental crust. Researchers also suggest that relatively buoyant asthenosphere has stayed attached to the base of the lithosphere and insulates the lithosphere from warmer parts of the asthenosphere. This special asthenosphere has been called the root of the craton.

In this essay, we discuss the Phanerozoic tectonics of the Midcontinent region of the United States. At first glance, the title may seem like an oxymoron, for the lack of topography and of penetrative structures gives the impression that this region has been tectonically stable during the Phanerozoic. However, while the Midcontinent has been *relatively* stable, it has not been *completely* stable. In fact, during the Phanerozoic, faults in the region have been reactivated, and broad regions of the surface have subsided to form sedimentary basins or have been uplifted to form domes or arches. Manifestations of Phanerozoic tectonism in the Midcontinent might not be as visually spectacular as that of North America's mountain belts, but tectonism has, nevertheless, occurred there.

22.7.2 Classes of Structures in the Midcontinent

To structural geologists raised on a diet of spectacular folds, faults, and deformation fabrics exposed in Phanerozoic orogens, the Midcontinent may seem, at first glance, to be structureless. But, in fact, the region contains four classes of tectonic structures: (1) epeirogenic structures, (2) Midcontinent fault-and-fold

TABLE 22.7.1 CATEGORIES OF CONTINENTAL CRUST

Active rift	A region where crust is currently undergoing horizontal stretching, so that crustal thicknesses are less than average crustal thicknesses. In active rifts, continental crust has a thickness of only 20–25 km, the crust has been diced up by normal faults, and volcanism occurs. Examples include the Basin and Range Province of the western USA, and the East African Rift.
Inactive rift	A belt of continental crust that underwent stretching, and became a narrow trough that filled with sediment, but never succeeded in breaking a continent in two. The Midcontinent rift of the central USA, and the Rhine Graben of Europe are examples.
Active orogen	A portion of the continental crust in which tectonism (faulting $\pm$ igneous activity $\pm$ uplift) currently takes place or has taken place in the recent past (Cenozoic). Such orogens tend to be linear belts, in that they are substantially longer than they are wide. Examples include the North American Cordillera and the European Alps.
Continental shelf	A belt fringing continents in which a portion of the continent has been submerged by the sea. Water depths over shelves are generally less than a few hundred meters. Continental shelves are underlain by passive-margin basins, which form subsequent to rifting, as a consequence of the subsidence of the stretched continental crust that bordered the rift. Sediment washed off the adjacent land buried the sinking crust. The stretching occurred during the rifting event and predated formation of a new mid-ocean ridge. Examples include the East Coast and Gulf coast of the USA.
Craton	A portion of a continent that has been relatively stable since the late Neoproterozoic (since at least about 800 Ma). This means that penetrative fabrics, regional metamorphism, and widespread igneous activity have not occurred in the craton during the Phanerozoic. The crust of a craton includes the eroded roots of Precambrian mountain belts.
Shield	A broad region, typically of low relief (though some shields have been uplifted and incised since the Mesozoic), where Precambrian crystalline rocks are exposed. In North America, a shield area (the Canadian Shield) is part of the craton. In South America, shield regions include cratons and Neoproterozoic orogens.
Continental platform	A broad region where Precambrian rocks (basement) have been covered by a veneer of unmetamorphosed Phanerozoic strata. Examples include the Midcontinent region of the USA, and large portions of northern Europe.
Inactive Phanerozoic orogen	Orogenic belts that were tectonically active in the Phanerozoic, but are not active today. Some inactive orogens, however, have been uplifted during the Cenozoic, so they are topographically high regions. Examples include the Appalachians of the USA and the Tasmanides of Australia.
Phanerozoic orogen	A general name for active Phanerozoic orogens and inactive Phanerozoic orogens, taken together.
Neoproterozoic orogen	Orogen active at the end of the Precambrian. Examples include the Pan-African/Brasiliano Orogens of Gondwana.

zones, (3) intragranular strain, and (4) regional joint systems. We'll describe these in succession.

EPEIROGENIC STRUCTURES When geologists first mapped the Midcontinent, they noted that Paleozoic strata of the region are nearly, but *not exactly* flat-lying. As a consequence, outcrop patterns of Paleozoic strata on a regional geologic map of the Midcontinent United States display distinctive bull's-eye patterns (Figure 22.7.2). In some examples, the youngest strata occupy the center of the bull's-eye, while in others, they occur in the outer ring. Bull's-eyes with the youngest strata in the center define intracratonic basins, in which strata are warped downward to form a bowl shape. Those with the youngest strata in the outer ring define intracratonic domes or, if elongate, intracratonic arches (by **"intracratonic"** we mean "within the craton"). As new data defining subsurface

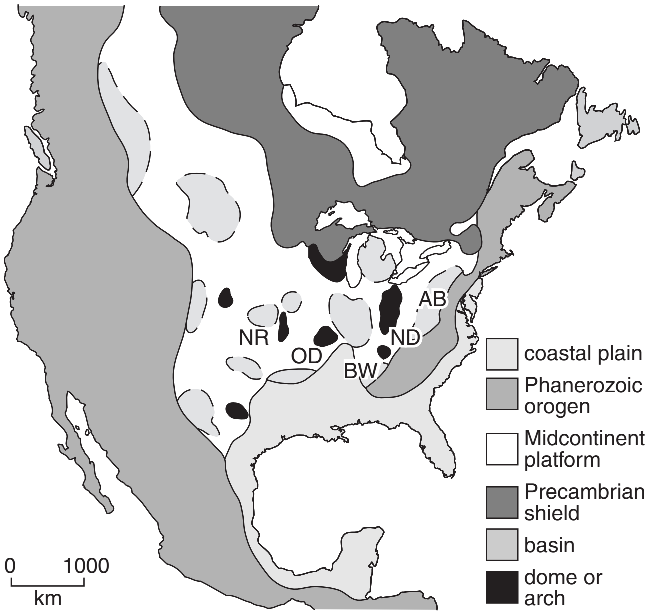

FIGURE 22.7.2 Regional tectonic map of North America showing the distribution of Phanerozoic orogenic belts, shield, continental-interior platform, and basins and domes.

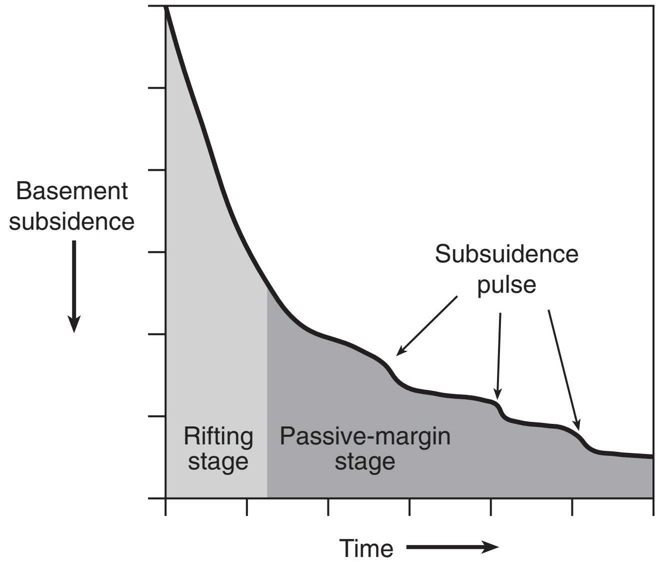

FIGURE 22.7.3 A graph illustrating how the rate of subsidence changes with time for an intracratonic basin. Rapid subsidence occurs at first, probably in association with rifting. Then, as the basin subsides due to cooling, the rate of subsidence gradually decreases. Pulses of rapid subsidence may occur during this time, perhaps due to the effects of orogeny along the continental margin. For example, an increase in stress may cause the lithosphere to weaken, so uncompensated loads beneath the basin may sink.

stratigraphy became available, primarily from correlation of well logs[3] and seismic-reflection profiles, researchers realized that stratigraphic formations thicken toward the center of a basin, and thin toward the center of a dome. Specifically, the Phanerozoic sedimentary cover of continental interior platforms thickens to about 5–7 km in basins, because that is where subsidence occurred most rapidly during deposition. In domes or arches, cover decreases in thickness to zero because these locations were emergent or were shallow shoals during deposition. A drop in sea level of the shallow oceans that covered the interior could expose the crest of a dome or arch, while the center of a basin could remain submerged. Thus, many more unconformities occur in the Phanerozoic section of domes and arches than in the interior of basins.

In effect, the lateral variations in sediment thickness that we observe in the Midcontinent indicate that there has been differential uplift and subsidence of regions of the Midcontinent during deposition. In other words, some regions went up while others went down as sediments were accumulating. Vertical movement affecting a broad region of crust is called **epeirogeny,** and the basins, domes, and arches that form as a result are **epeirogenic structures.** Significantly, the general position of basins and domes in the Midcontinent has remained fixed through the Phanerozoic. For example, the Illinois Basin and Michigan Basin have always been basins while the Ozark Dome and the Cincinnati Arch have been relatively high since their initial formation in the Late Proterozoic or Early Cambrian. Thus, these epeirogenic structures represent permanent features of continental-interior lithosphere.

Lack of stratigraphic evidence, either due to nondeposition or erosion, makes it impossible to constrain the rate of uplift of arches and domes precisely, but we can constrain rates of epeirogenic movement in basins by studying subsidence curves, which are graphs that define the rate at which the basement-cover contact at the floor of the basin moved down through time. Geologists generate subsidence curves by plotting sedimentary thickness as a function of time, after taking into account the affect of compaction (Figure 22.7.3). Subsidence curves for intracratonic basins demonstrate

[3]When drilling for oil, exploration companies use a rotating drill bit, which penetrates into the earth by grinding the rock into a mixture of powder and small chips called "cuttings." Circulating drilling "mud," pumped down into the hole, flushes the cuttings out of the hole. Since this process does not yield an intact drill core, the only way to determine the precise depth at which specific rock units lie in the subsurface is to lower instruments down the hole to record parameters such as electrical resistively and gamma-ray production, parameters that correlate with rock type. A record of resistivity versus depth, or gamma-radiation versus depth, is called a well log.

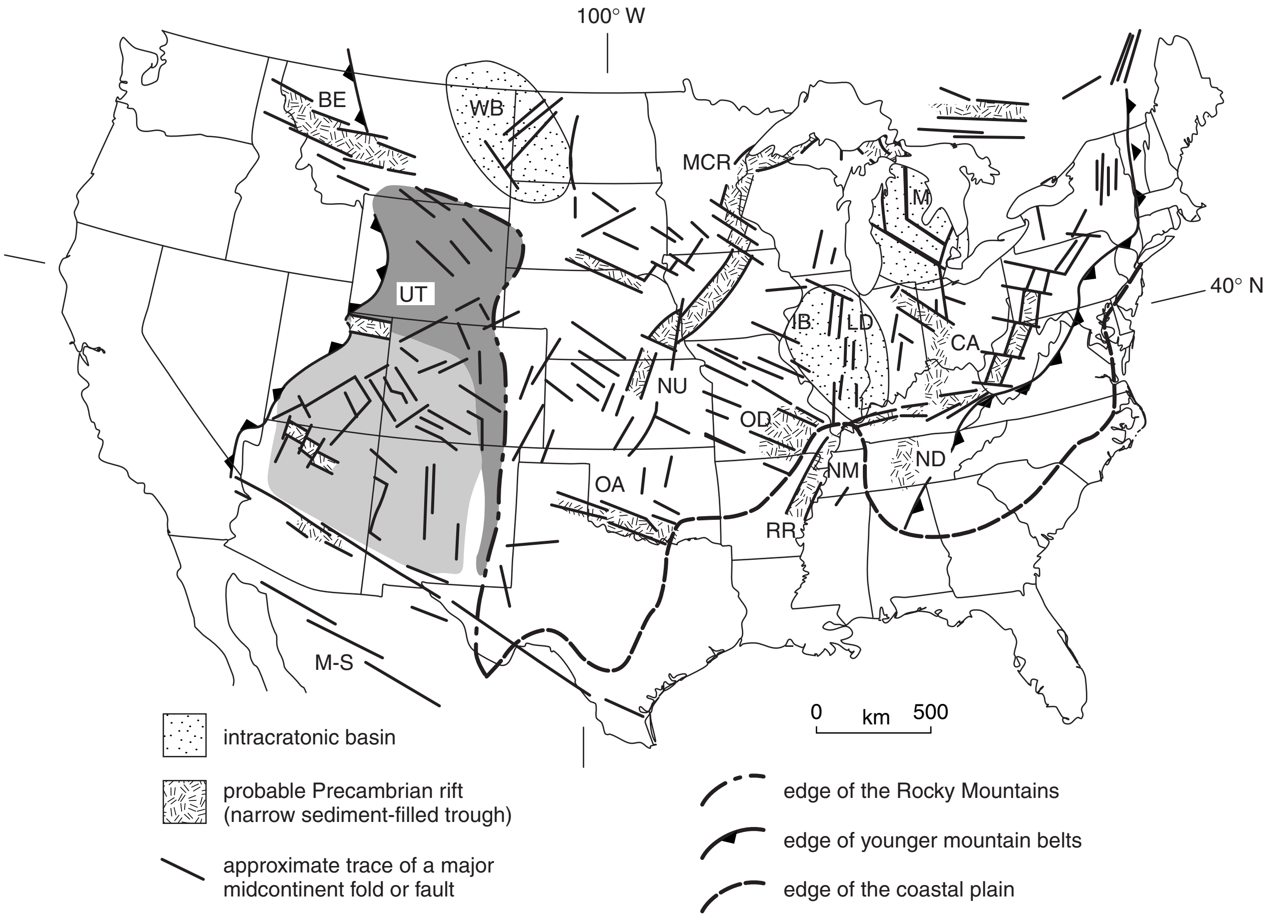

FIGURE 22.7.4 Map of the United States showing the distribution of Midcontinent fault-and-fold zones and of documented intracratonic rifts. BE = Beltian embayment, UT = Uinta trough, MCR = Midcontinent Rift, OA = Southern Oklahoma aulacogen, RR = Reelfoot Rift, LD = La Salle deformation belt, WB = Williston Basin, IB = Illinois Basin, MB = Michigan Basin, ND = Nashville dome, CA = Cincinnati arch, M-S = Mojave-Sonora megashear.

that basins have subsided overall through most of the Phanerozoic, probably due to slow cooling of lithosphere beneath the basins, but emphasize that the rate of subsidence for a basin varies with time. For example, pulses of rapid subsidence occurred in the Michigan and Illinois Basins during Ordovician, Late Devonian–Mississippian, and Pennsylvanian–Permian time. Notably, the timing of some, but not all, epeirogenic movements in the Midcontinent roughly corresponds with the timing of major orogenic events along the continental margin. For example, the time of Late Paleozoic subsidence in the Midcontinent U.S. basins roughly corresponds with the time of the Alleghanian Orogeny (the collision of Africa with the eastern margin of North America).

MIDCONTINENT FAULT-AND-FOLD ZONES Bedrock geology of much of the Midcontinent has been obscured by Pleistocene glacial deposits and/or thick soils. In the isolated exposures that do occur, geologists have found occurrences of folds and faults. The regional distribution of such structures has slowly emerged from studies of well logs (compiled on structure-contour and isopach maps), potential-field data (compiled on gravity- and magnetic-anomaly maps), and seismic-reflection profiles. Abrupt steps and ridges on structure-contour maps or isopach maps, linear anomalies on potential field maps, abrupt bends or breaks in reflectors on seismic-reflection profiles—along with outcrop data—reveal that the Phanerozoic strata of the Midcontinent United States has been disrupted locally by distinct belts of deformation that we refer to here as Midcontinent fault-and-fold zones (Figure 22.7.4).

Individual fault-and-fold zones range in size from only a few hundred meters wide and several kilometers long, to 100 km wide and 500 km long. The north-northwest trending La Salle fault-and-fold zone of

Illinois serves as an example of a larger zone—it effectively bisects the Illinois Basin (Figure 22.7.5). Larger zones include numerous non-coplanar faults that range in length from <5 km to as much as 50 km. At their tips, these faults overlap with one another in a relay fashion. Locally, a band of *en echelon* subsidiary fault segments borders the trace of principal faults. In the upper few kilometers of the crust, major faults of a Midcontinent fault-and-fold zone dip steeply and divide upwards into numerous splays. The resulting array of faults resembles that of a flower structure. At depth, major faults decrease in dip (i.e., some faults appear to be listric) and penetrate basement (Figure 22.7.6). Some, but not all, faults clearly border narrow rift basins that contain anomalously thick sequences of sediments and volcanics. The largest of these intracratonic rifts, the Midcontinent Rift, consists of two principal arms, one running from Lake Superior into Kansas and the other running diagonally across Michigan (Figure 22.7.4). Faults along these rifts initiated as normal faults, but later reactivated as thrust or strike-slip faults.

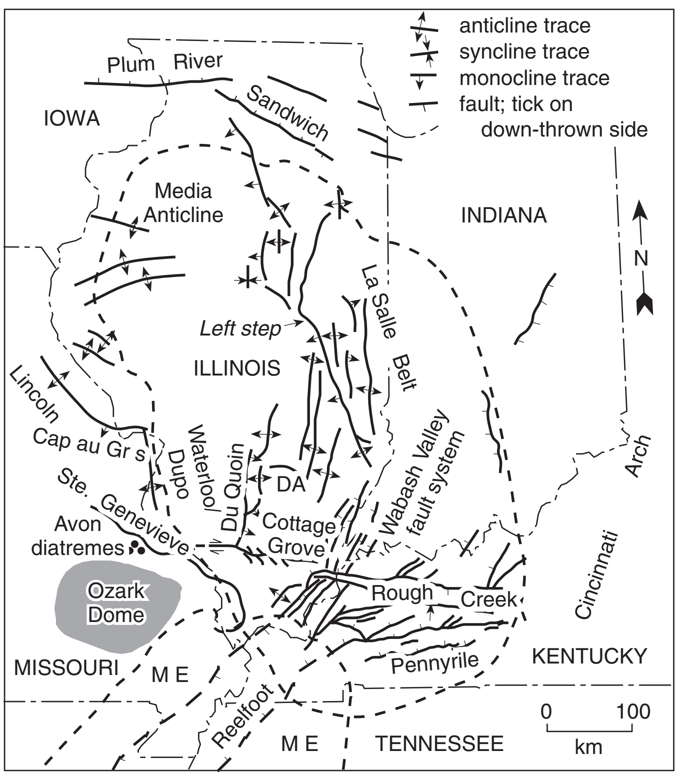

FIGURE 22.7.5 Structure map of the Illinois Basin region showing the map traces of Midcontinent fault-and-fold zones. Note that the Cottage Grove Fault is bordered by short *en echelon* faults, indicative of strike-slip displacement.

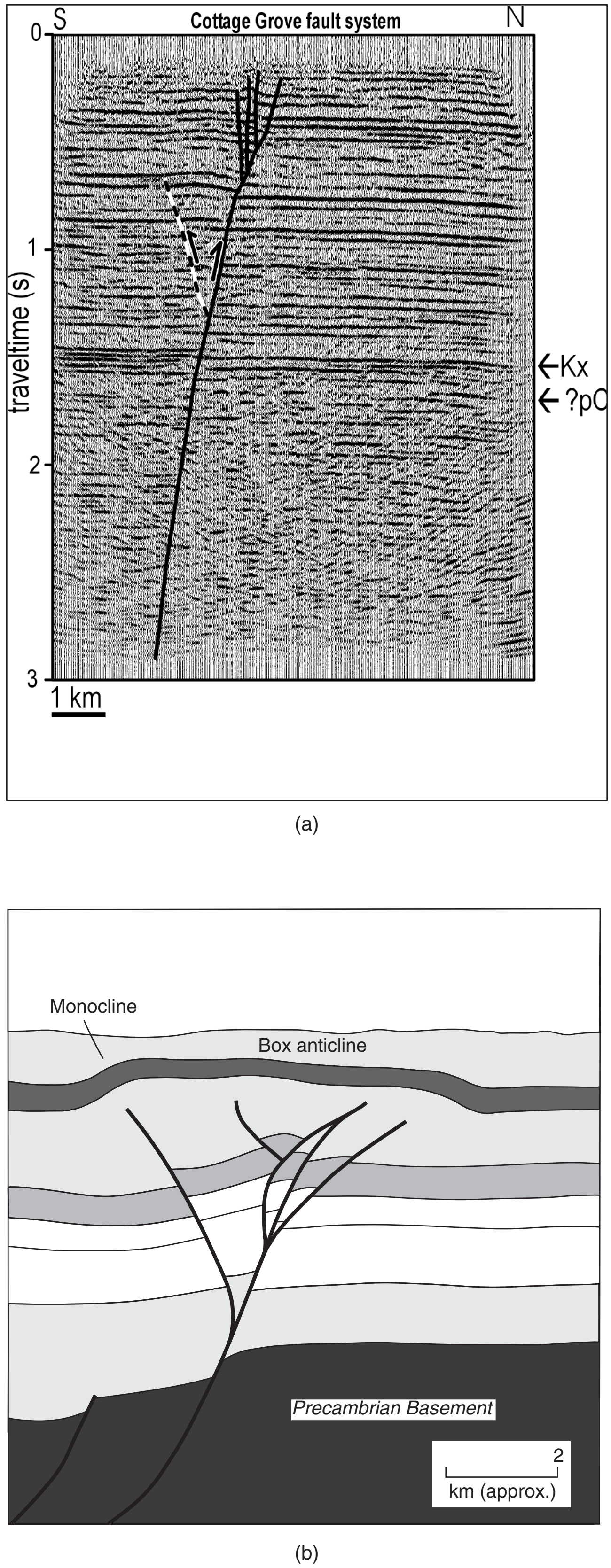

FIGURE 22.7.6 Cross-sectional structure of a Midcontinent fault zone. (a) Seismic-reflection profile of the Cottage Grove fault system; (b) Schematic cross section showing the principal features of a Midcontinent fault zone.

Cross sections indicate that major faults in the Midcontinent locally have vertical throws of as much as 2 km, but more typically throws are less—no more than tens to hundreds of meters. Strike-slip components of displacement across continental-interior faults are difficult to ascertain because of lack of exposed shear-sense indicators on fault surfaces and the lack of recognizable offset markers. But geologists have found strike-slip lineations on faults exposed in coal mines, and the *en echelon* map pattern of subsidiary faults adjacent to larger faults resembles the *en echelon* faulting adjacent to continental strike-slip faults. Such features suggest that a component of strike-slip motion has occurred on some Midcontinent faults. Where such motion has occurred, the faults can be considered to be oblique-slip faults. The occurrence of oblique slip, in turn, suggests that transpression or transtension has taken place in some Midcontinent fault-and-fold zones.

In general, folds of the continental interior are monoclinal in profile, meaning that they have one steeply dipping limb and one very shallowly dipping to subhorizontal limb. Locally, oppositely facing monoclinal folds form back-to-back, creating "box anticlines" (Figure 22.7.6b). Though geologists have not yet obtained many clear images of Midcontinent folds at depth, several studies document that folds lie above steeply dipping faults and that structural relief on folds increases with depth.

Detailed study of spatial variations in the thickness and facies of a stratigraphic unit relative to a fault-and-fold zone, documentation of the timing of unconformity formation, as well as documentation of local slump-related deformation, permits determination of when the fold-and-fault zone was tectonically active. Such timing constraints suggest that the structures, in general, became active during more than one event in the Phanerozoic. Activity appears to have been particularly intense during times of orogenic activity along the continental margin, but occurred at other times as well. The most significant reactivation occurred during the late Paleozoic, at the same time as the Alleghanian Orogeny. This event triggered reactivation of faults across the entire interior platform. Fault reactivation in the region that now lies within the Rocky Mountains yielded localized uplifts (now eroded) that have come to be known as the Ancestral Rockies. Some authors now use this term for Late Paleozoic uplifts associated with faulting across the width of North America (Figure 22.7.7).

Midcontinent fault-and-fold zones do not have random orientations, but rather display dominant trends over broad regions of the craton. As indicated by the map in Figure 22.7.4, most of the zones either trend north-south to northeast-southwest, or east-west to northwest-southeast, and thereby outline rectilinear blocks of crust. Along-strike linkage of fault-and-fold zones seems to define transcratonic belts of tectonic reactivation, which localize seismicity today.

Of note, Midcontinent folds closely resemble Laramide monoclines of the Colorado Plateau, and Laramide basement-cored uplifts of the U.S. Rocky Mountain region, though

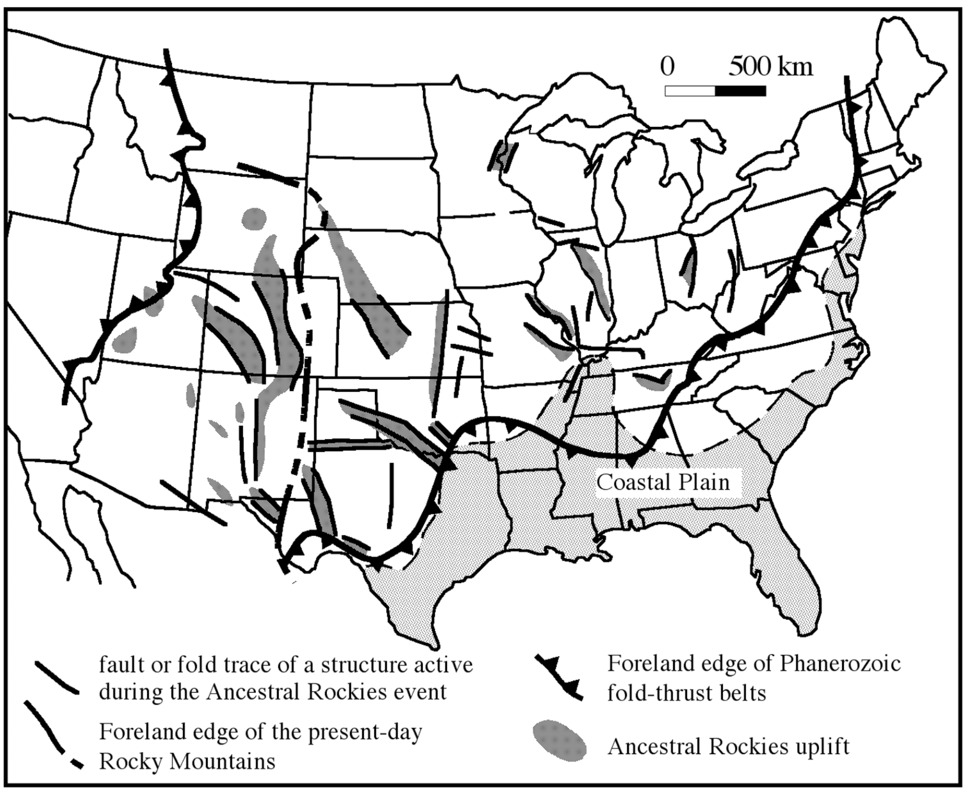

FIGURE 22.7.7 Late Paleozoic uplifts of the United States. The large ones west of the Rocky Mountain front have traditionally been referred to as the Ancestral Rockies.

at a smaller scale. This is not surprising because, as we noted earlier, the Colorado Plateau and Rocky Mountain region were once part of the craton's interior platform. These regions differ from the Midcontinent only in that they were uplifted and deformed during the Late Mesozoic–Early Cenozoic Laramide Orogeny. During this event, slip on some faults of the region exceeded 15 km, an order of magnitude more than occurred during Paleozoic reactivation of faults in the Midcontinent. Because of substantial Laramide and younger uplift, and because of the dry climate of the Plateau, fault-and-fold zones of the Plateau are brazenly displayed. But keep in mind that, if the Midcontinent region were stripped of its glacial sedimentary blanket and its prairie soils, it would look much like the Colorado Plateau.

INTRAGRANULAR STRAIN AND REGIONAL JOINTING When you traverse a large fold in the Appalachian fold-thrust belt, you will find that outcrops of argillaceous (clay-rich) sandstones and limestones contain well-developed spaced cleavage to slaty cleavage. But, if you walk across a large fold in the Midcontinent (or on the Colorado Plateau), you will find a distinct lack of cleavage. Layer-parallel shortening strains sufficient to form a regional cleavage apparently did not develop in the Midcontinent. Microstructural studies, however, indicate that subtle layer-parallel shortening did, in fact, develop in Midcontinent strata. This strain is manifested by the development of calcite twinning, a type of intragranular, crystal-plastic strain that can be seen only with a microscope and has developed in limestone.

Calcite twinning forms under the relatively low pressure and temperature conditions that are characteristic of the uppermost crust. Regional studies of calcite twinning in Midcontinent limestones indicate that the maximum shortening direction remains fairly constant over broad regions and trends roughly perpendicular to orogenic fronts, though more complex shortening patterns occur in the vicinity of Midcontinent fault-and-fold zones (Figure 22.7.8). Notably, strain and differential stress magnitudes in the eastern Midcontinent decrease progressively from the Appalachian-Ouachita orogenic front to the interior—strain magnitudes are 6% at the front but decrease to 0.5% in the interior,

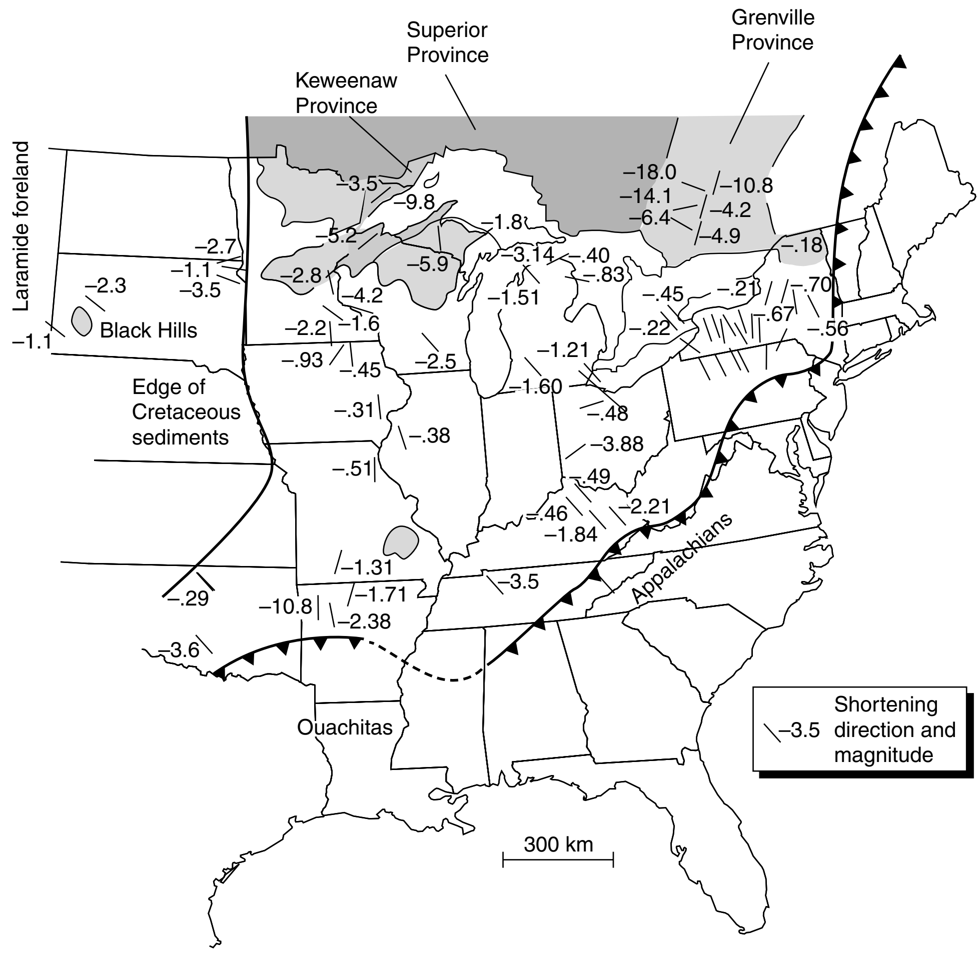

FIGURE 22.7.8 Calcite twinning strains in eastern North America. Regional tectonic provinces are labeled, except for the Paleozoic cover sequence inland from the Appalachian-Ouachita thrust front (bold, toothed line). Strains are presented by orientation (short lines) and magnitude in percent (negative is shortening); typically, maximum shortening is horizontal and perpendicular to the Appalachian-Ouachita thrust front. Twinning data from other tectonic provinces show patterns that are unrelated to Paleozoic deformation.

with associated differential stresses exponentially decreasing from ~100 MPa to less than 20 MPa. This pattern suggests that calcite twinning in strata of the eastern Midcontinent formed during the Alleghanian Orogeny, the Late Paleozoic collision of Africa with North America. In the western part of the Midcontinent, twinning strains also generally lie in the range of 0.5% and 3%, and the direction of maximum shortening trends roughly at right angles to the Rocky Mountain front. This geometry implies that layer-parallel shortening in strata of the western Midcontinent developed in association with compression accompanying the Sevier and Laramide Orogenies.

No one has yet compiled joint-trend data for the entire Midcontinent, but the literature does provide data from numerous local studies. Joint frequency diagrams suggest that there are two dominant joint sets (one trending generally northwest and one trending generally northeast) and two less prominent sets (one trending east-west and one trending north-south) in the Devonian strata of northern Michigan. Similar, but not exactly identical, trends have been documented in Ohio, Indiana, Illinois, and Wisconsin. Taken together, regional studies suggest that systematic vertical joint sets do occur in platform strata of the Midcontinent, and that in general there are east-west sets, northwest sets, north-south sets, and north-east sets, but that orientations change across regions and that different sets dominate in different locations. The origin of this jointing remains enigmatic.

22.7.3 Some Causes of Epeirogeny

Over the years, geologists and geophysicists have proposed many mechanisms to explain continental interior epeirogeny. The candidates that may explain epeirogeny will be briefly discussed and are illustrated in Figure 22.7.9.

- *Cooling of Unsuccessful Rifts.* Intracratonic basins may form because of thermal contraction due to cooling of unsuccessful rifts that opened during the Late Proterozoic. When active extension ceased, these rifts cooled and subsided, much like the rifts that underlie passive margins. Such epeirogeny continued through the Phanerozoic at ever-decreasing rates.
- *Variations in Asthenosphere Temperature.* Modern seismic tomography studies of the Earth's interior demonstrate that the Earth's mantle is thermally heterogeneous. As lithosphere drifts over this heterogeneous asthenosphere, it conceivably warms when crossing hot asthenosphere and cools when crossing cool asthenosphere. These temperature changes could cause isostatic uplift (when warmed) or subsidence (when cooled) of broad regions of the lithosphere.
- *Changes in State of Stress.* Changes in stress state in the lithosphere may cause epeirogenic movement in many ways. For example, as differential stress increases, plastic deformation occurs more rapidly, so that the viscosity of the lithosphere effectively decreases. Thus, an increase in differential stress weakens the lithosphere. If this were to happen, denser masses in the crust (e.g., a lens of mafic igneous rock below a rift), which were previously supported by the flexural strength of the lithosphere, would sink, whereas less dense masses (e.g., a granite pluton) would rise. Thus, differential epeirogenic movements may be localized by preexisting heterogeneities of the crust, which is set free to move in an attempt to attain isostatic equilibrium by the weakening of the lithosphere that accompanies an increase in differential stress. Some geologists have suggested that changes in horizontal stress magnitude may also cause epeirogeny by buckling the lithosphere, or by amplifying existing depressions (basins) or rises (arches).
- *Flexural Response to a Load.* Creation of a large load, such as a volcano or a stack of thrust sheets, results in flexural loading on the surface of the continent, and thus bending down of the continent's surface. Flexural loading due to emplacement of thrust sheets leads to the development of asymmetric sedimentary basins, called foreland basins, on the craton side of fold-thrust belts. In addition to causing a depression to form, the levering effect of the loaded lithosphere may cause an uplift, or outer swell, to form on the cratonic-interior of the depression. This effect may have caused Midcontinent uplifts, like the Cincinnati arch, to rise in response to loading the continental margin by thrust sheets of the Appalachian orogen.
- *Block Tilting.* As noted earlier, Midcontinent fault-and-fold zones divide the upper crust into fault-bounded blocks. Changes in the stress state in the continental interior may cause tilting of regional-scale, fault-bounded blocks of continental crust relative to one another. These could cause uplift or subsidence of the corners and edges of blocks.
- *Changes in Crustal-to-Lithosphere Mantle Thickness Ratio.* Continental elevation is controlled, regionally, by isostasy. Since crust and mantle do not have the same density, any phenomenon that causes a change in the proportion of crust to lithospheric mantle in a column from the surface of the Earth down to the level of isostatic compensation

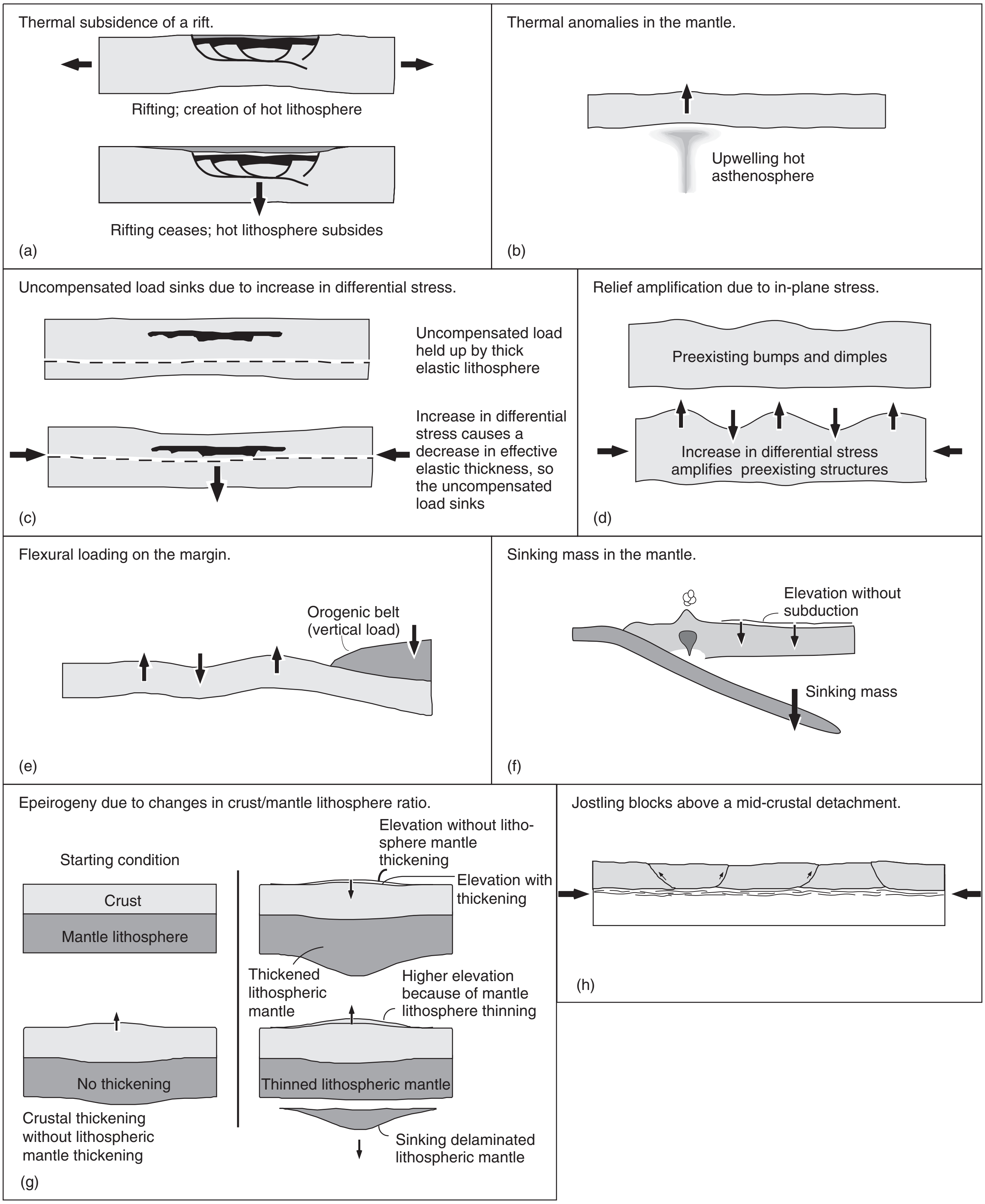

FIGURE 22.7.9 Models of epeirogeny. (a) Thermal cooling over an unsuccessful rift (before and after); (b) Uplift related to thermal anomalies in the mantle; (c) Vertical movement of an uncompensated load due to changes in the elastic thickness of the lithosphere; (d) Amplification of preexisting bumps and dimples due to in-plane stress; (e) Flexural loading of a lithospheric margin; (f) Epeirogeny related to subduction; (g) Epeirogeny due to changes in the ratio of crustal thickness to lithosphere mantle thickness; (h) Epeirogeny due to tilting of regional fault-bounded blocks.

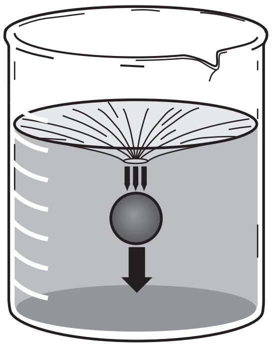

FIGURE 22.7.10 Schematic drawing showing the sinking ball model for epeirogeny caused by subduction (or dynamic topography). As the ball sinks through the honey (shaded area), the thin plastic film on the surface of the honey is pulled down.

has the potential to cause a change in the elevation of the continent's surface. For example, thickening of the crust, perhaps due to plastic strain in response to tectonic compression relative to the lithospheric mantle, would cause a rise in elevation; decreasing the thickness of the lithospheric mantle in response to delamination (separation of lithospheric mantle from the base of the plate) could also cause a rise in crustal elevation; adding basalt to the base of the crust (a process called underplating) would thereby thicken the crust and could cause uplift.

- *Subduction.* Subduction of oceanic lithosphere can cause epeirogenic movement because, as the subducted plate sinks, it pulls down the overlying continent. To picture this phenomenon, imagine a bucket filled with honey (representing the asthenosphere), in which an iron ball (representing oceanic lithosphere) has been suspended just below the surface (Figure 22.7.10). Now, place a film of plastic wrap (representing the continental lithosphere) over the top of the honey, and then release the ball. As the ball sinks (representing subduction of the oceanic lithosphere), the plastic wrap is pulled down. This downward motion is epeirogenic subsidence.

22.7.4 Speculations on Midcontinent Fault-and-Fold Zones

As noted above, stratigraphic evidence demonstrates that Midcontinent fault-and-fold zones were reactivated multiple times during the Phanerozoic. But how did the zones originate in the first place? Did the faults initiate during the Phanerozoic by brittle rupturing of intact crust in response to compression caused by orogeny along the continental margin? If so, then the faults started out as reverse or transpressional faults. Or did the structures form earlier in Earth history? We favor the second proposal, and suggest that Midcontinent fault-and-fold zones initiated as normal faults during episodes of Proterozoic extension. Thus, displacements in the zones during the Phanerozoic represent fault reactivation. We base this statement on the observation that Midcontinent fault-and-fold zones have the same trends as Proterozoic rift basins and dikes in the United States. They are not systematically oriented perpendicular to the shortening directions of marginal orogens.

If the above hypothesis is correct, then the fault-and-fold zones of the Midcontinent, as well as of the Rocky Mountains and Colorado Plateau, are relicts of unsuccessful Proterozoic rifting. Once formed, they remained as long-lived weaknesses in the crust, available for reactivation during the Phanerozoic, when the interior underwent slight regional strain. A reverse component of displacement occurred on the faults if reactivation was caused by regional shortening, whereas a normal sense of displacement occurred if reactivation was caused by regional extension. On faults that were not perpendicular to shortening or extension directions, transpression or transtension led to a component of strike-slip motion on faults. Regarldless of slip direction, displacement on the faults generated fault-propagation folds in overlying strata. Note that reverse or transpressional motion represents inversion of Proterozoic extensional faults, in that reactivation represents reversal of slip on the faults. In cases where the faults bounded preserved rift basins, the inversion led to thrusting of the rift's contents up and over the rift's margin. But rift basins do not occur in all fault-and-fold zones, either because Proterozoic displacements were too small to create a basin, or because the basin was eroded away during latest Proterozoic rifting, before Phanerozoic strata were deposited.

The hypothesis that Midcontinent fault-and-fold zones are reactivated Proterozoic normal faults is appealing because it explains how these structures could have formed with the orientations that they have, and how they formed without the development of regional cleavage. Zone orientation simply reflects the trends of preexisting Proterozoic normal faults, not the orientation of a regional stress field during the Phanerozoic. The lack of regional cleavage reflects the fact that development of these structures is not associated with significant shortening above detachments within the Phanerozoic section. This hypothesis also explains the timing of movement—faults were reactivated primarily during marginal orogenies or rifting events, when displacement of the continental margin caused a slight strain in the interior, and this strain was accommodated by movement on faults. Perhaps a way to envision Phanerozoic Midcontinent faulting and folding is to think of the upper crust in the craton as a mosaic of rigid fault-bounded blocks that jostle relative to one another in response to changes in the stress state of the continental interior (Figure 22.7.11). Depending on the geometry of the stress during a given time period, blocks may move slightly apart, move slightly together, or move laterally relative to one another. Movements tend to be transpressional or transtensional, for the belts are not oriented appropriately for thrust, reverse, or strike-slip faulting alone to occur.

The greatest amount of movement in Midcontinent fault-and-fold zones occurred in Late Paleozoic time, when Africa collided on the east, South America collided on the south, and a subduction zone had formed along the southwest. This pulse resulted in the formation of the Ancestral Rockies of the Rocky Mountains Province as well in the kilometer-scale displacements in fault-and-fold zones across the Midcontinent (Figure 22.7.7). The intracratonic strain that formed at this time is significantly less than that in Asia during the Cenozoic collision of India with Asia. The contrast in continental-interior response to collision probably reflects the respective strengths of these two continents. The interior of Asia consists of weak continental crust of the Altaids Orogen, while the interior of the United States is a strong craton.

While major movements appear to have accompanied major marginal orogenies, movement on these faults can occur during nonorogenic times as well. For example, historic intraplate earthquakes (earthquakes occurring in a plate interior, away from plate boundaries) at New Madrid, Missouri, result from movements at the intersection of two Midcontinent fault-and-fold zones. This movement may be a response to the ambient stress in the continental lithosphere, caused by ridge-push force and/or basal traction, or to stress resulting from epeirogenic movements.

22.7.5 Closing Remarks

The speculative tone used in this essay emphasizes that geologists need to obtain more data on structures in cratonic interiors before we can confidently explain them and assess their significance. However, it has become increasingly clear that cratonic interiors were not tectonically dead during the Phanerozoic. Rather, they were sensitive recorders of plate interactions, which may have caused jostling of upper-crustal blocks. Although we have focused on examples of structures in the Midcontinent of the United States,

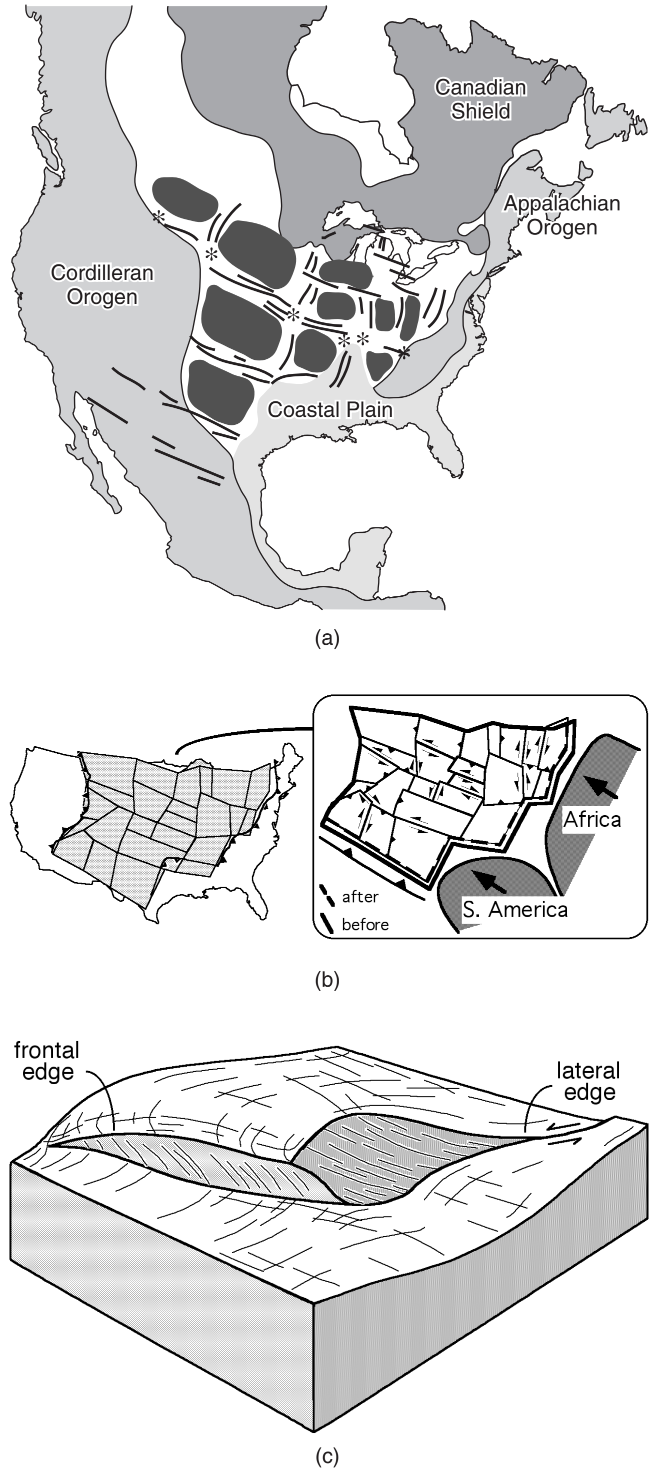

FIGURE 22.7.11 (a) Map showing block model of intracratonic tectonism in the United States. Stars indicate seismically active regions. The diagram in (b) illustrates how thrust and strike-slip motions can be reactivated in response to regional compression, while the diagram in (c) illustrates how uplift along a reverse fault leads to oblique slip along the side of the block.

keep in mind that these structures can be viewed as type examples for continental-interior platforms throughout the world. Finally, it is important to emphasize once more that the dramatic basement-cored uplifts that developed in the Rocky Mountains, as well as in the Sierra Pampeanas of Argentina, and the Tien Shan of southern Asia, are similar in style to those of the U.S. Midcontinent; all may have been formed by reactivation of preexisting faults.

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