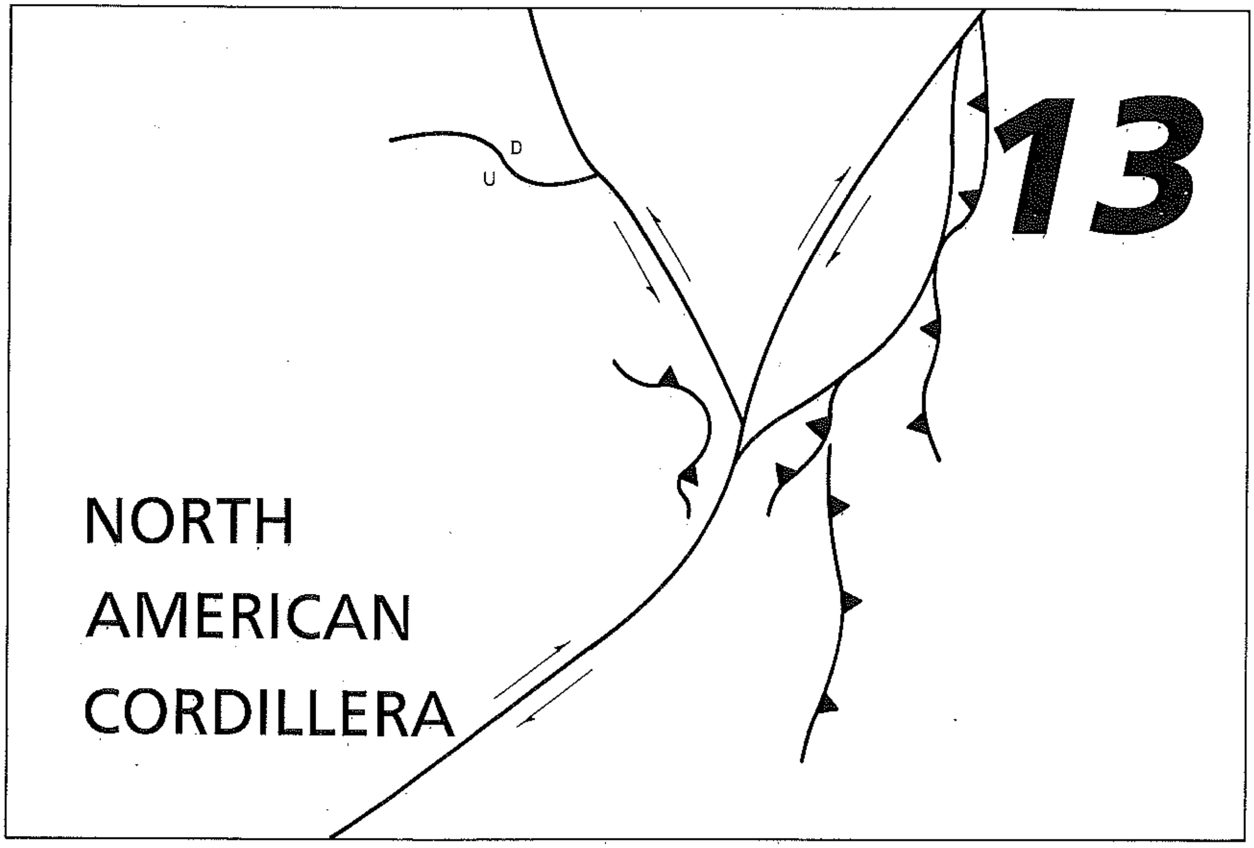

13

NORTH AMERICAN CORDILLERA

INTRODUCTION

The great *Cordilleran mountain belt* (Figure 13-1) composes the deformed western border of the North American continent, extending from Alaska to Central America, with further extensions of deformation into northeastern Asia, the Caribbean, and the Andes of South America. In contrast with the Paleozoic Appalachian mountain belt discussed in the preceding chapter, the Cordilleran mountain belt is undergoing present-day deformation over its entire length and over much of its width, as demonstrated by current seismicity, volcanism, and mountainous topography. The Cordilleran mountain belt has been a locus of continental-margin deformation, at least sporadically, since the late Paleozoic, about 350 m.y. Furthermore, it has never experienced the suturing of megacontinental masses in the manner of the Appalachian-Caledonide chain and parts of the Alpine-Himalayan system; it has always bordered a major "Pacific" ocean. The Cordilleran belt displays an orogenic style in marked contrast with the Appalachian style introduced in Chapter 12.

One of the most immediately instructive aspects of the Cordilleran belt is the fact that the presently active mountain belt does not represent a single plate boundary but involves deformation between the North American plate and the Pacific, Gorda, Cocos, and Caribbean plates. From place to place, the mountain belt is characterized by active compressive, strike-slip, and extensional deformation. The relative motion between the major North American and Pacific plates is right-lateral strike slip, approximately parallel to the continental margin in Canada, California, and northern Mexico (Fig. 13-1). Therefore, active strike-slip faults, such as the San Andreas in California and the Queen Charlotte offshore of western Canada, exist where the North American and Pacific plates are in

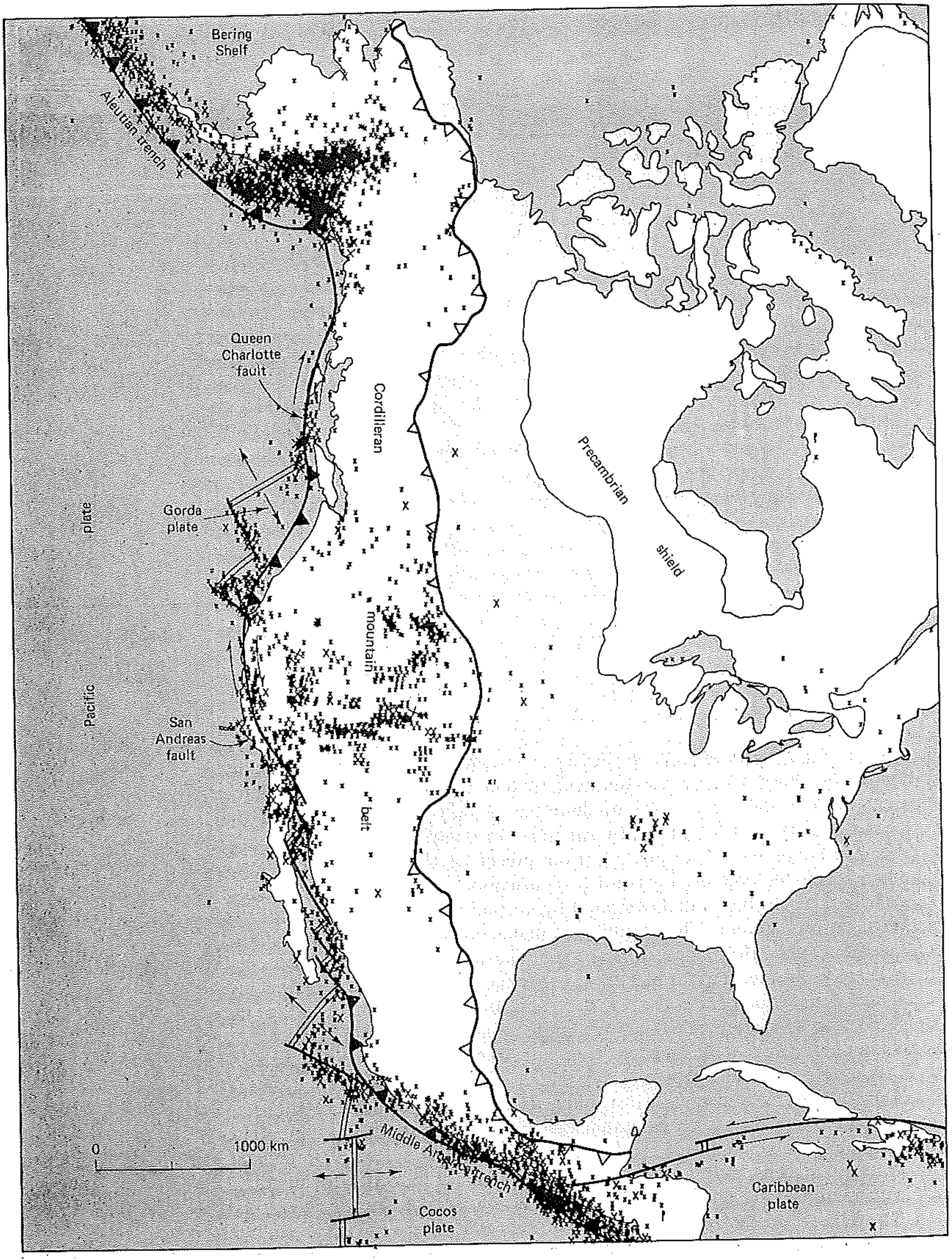

FIGURE 13–1 The Cordilleran mountain belt of western North America and its present-day seismicity and plate configuration. (Seismicity courtesy of U.S. Geological Survey.)

contact. The same North American–Pacific plate boundary becomes compressive, marked by subduction and arc volcanism, where the continental margin bends perpendicular to the plate motion in southern Alaska and the eastern Aleutians. The plate boundary leaves the North American continent with the Aleutian island arc, which gradually bends to parallelism with the plate motion as it approaches the Kamchatka Peninsula of Siberia 2000 km to the west. The Bering Shelf of western Alaska is the only part of the western continental margin of North America that is not marked by present-day plate-boundary tectonics. In summary, important variation in the style of present-day Cordilleran tectonics between right-lateral strike slip and subduction primarily reflects the angle between the continental margin and the relative-motion vector.

A second major cause of present-day tectonic variation along the length of the Cordillera is several oceanic plates that intervene between the Pacific plate and the Cordillera: the Gorda and Cocos plates (Fig. 13-1). These smaller plates give rise to subduction and volcanic-arc tectonics in the northwestern United States, southern Mexico, and Central America. South of the strike-slip fault zone in central Guatemala, the Cordillera lie on the Caribbean plate rather than the North American plate.

The history of interaction between North America and the Pacific and intervening plates has led to a similarly complex variety of tectonic behavior along the Cordillera in the Cenozoic, as will be discussed later in this chapter. For the moment, the important principle illustrated by this brief tour of present-day plate interactions is that a great mountain belt of continental or intercontinental extent should not, in general, be equated with a single plate boundary even at a single time, but rather a whole system of plate boundaries, often localized at a continent-ocean interface. This fact could equally well be illustrated with the present-day Andes or the Alpine-Himalayan chain (Fig. 1-30). The continent-ocean interface is inherently unstable, making it a common site of plate interactions.

We can expect equally complex plate interactions to have marked the tectonic history of the Cordillera in the last 350 m.y. Much of this history is yet to be deciphered over the 12,000-km length of the chain. Our purpose in this chapter is to introduce only the most-throughgoing and best-understood themes in the regional structural geology and tectonic history of the Cordillera, particularly in the United States and adjacent Canada and Mexico.

Our methods of approaching the tectonics of the Cordillera will initially be similar to those used in the Appalachians. We begin from the North American continental interior and move into the eastern marginal zones of deformation, which involve rocks of strong North American affinities. As we move west, we eventually encounter rocks that have strong oceanic affinities or show signs of being highly allochthonous. This is all very similar to the Appalachians in some ways. In contrast, other aspects of orogenic belts that have not been prominent in our discussion of Appalachian tectonics, such as strike-slip faulting, voluminous outpouring and intrusion of magma, plateau uplift, and normal faulting, are found to play a major role in the Cordillera.

CRATONIC FORELAND

The eastern edge of strong deformation in the Cordilleran mountain belt is generally close to the eastern edge of thick Paleozoic sedimentation, although there are important exceptions to be emphasized later. Paleozoic sedimentary thicknesses east of the eastern fold-and-thrust belt (Fig. 13-2) are generally 2 km or less, typical of cratonic stratigraphic sections (Fig. 1-8). For example, this

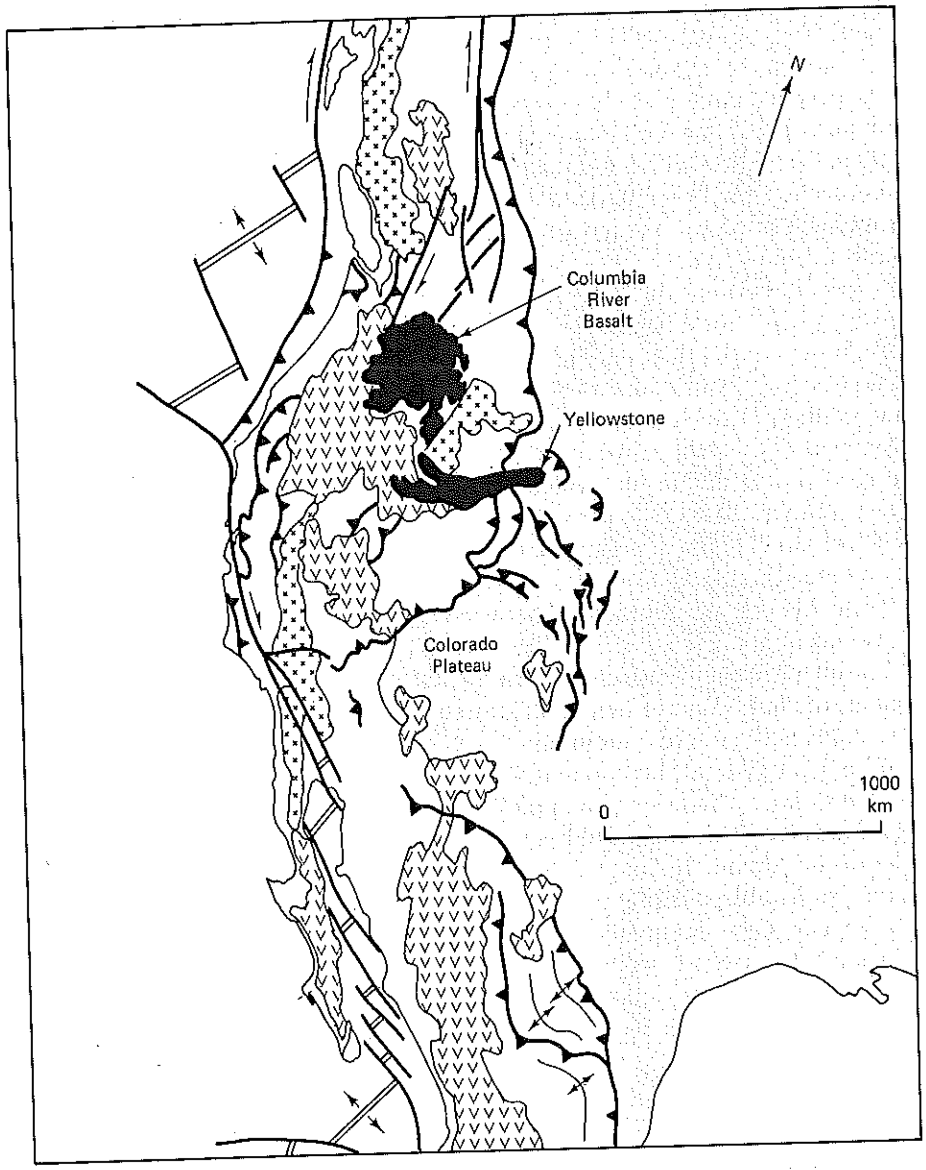

FIGURE 13–2 Major tectonic features of the Cordilleran mountain belt from southern Canada to northern Mexico. The batholith belt is shown in the x pattern and Cenozoic volcanic rocks are shown in the v pattern. The eastern edge of the mountain belt is the east-vergent fold-and-thrust belt.

cratonic stratigraphy is exposed in the Grand Canyon of the Colorado River, where the Cambrian through Permian sequence is only about 1200 m thick (Fig. 13-3). The overlying Triassic and Jurassic sediments that are exposed in nearby basins to the north increase the total foreland thickness by only another kilometer. Similar cratonic thicknesses are typical of most areas east of the fold-and-thrust belt.

In contrast, if we turn west from the Grand Canyon and travel into the core of the Cordilleran orogenic belt, we find that the Paleozoic section passes continuously into a basin of geosynclinal proportions, as shown in Figure 13-4, which is a restored stratigraphic cross section. The Permian and older section near Las Vegas is about 8 km thick, in contrast with 1200 m at the Grand Canyon (Fig. 13-3) only 250 km to the east. This thickening is quite analogous to the thickening of sediments into the Appalachian mountain belt discussed in the last chapter (Fig. 12-5). The increased thickness is accomplished in three ways: (1) greater rate of

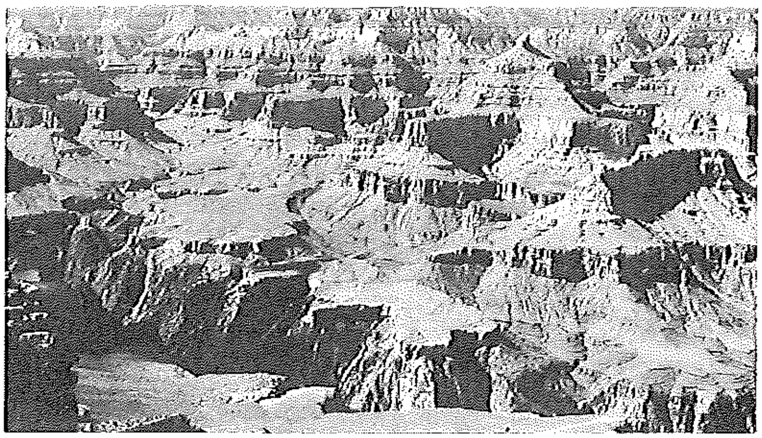

FIGURE 13–3 The Grand Canyon of the Colorado River, Arizona, exposing the thin (1200 m) Paleozoic cratonic stratigraphy of the Colorado Plateau, east of the Cordilleran geosyncline. Precambrian basement and Proterozoic sedimentary rocks are exposed in the canyon bottom.

subsidence, (2) more-complete stratigraphic sequence with less time missing in disconformities, and (3) older rocks at the base of the section. Most of the thick Cordilleran section represents a carbonate platform at the edge of the craton, dominated by shallow-water limestone, dolomite, and quartzites. This carbonate-quartzite sequence is analogous to the Cambrian–Lower Ordovician carbonate-quartzite sequence of the Appalachian foreland, which preceded the Taconic orogeny, and might be analogous to the present-day Florida, Bahama, and Yucatan platforms on the tropical edges of North America.

Underlying the Paleozoic carbonate bank is a very thick sequence of Upper Proterozoic clastic sediments of both deep- and shallow-water deposition, including minor basaltic volcanic rocks; they are called the *Windermere Series* in Canada. This section is in many ways analogous to the Upper Precambrian clastic section of the southern Appalachians that records the initial rifting of the Proto-

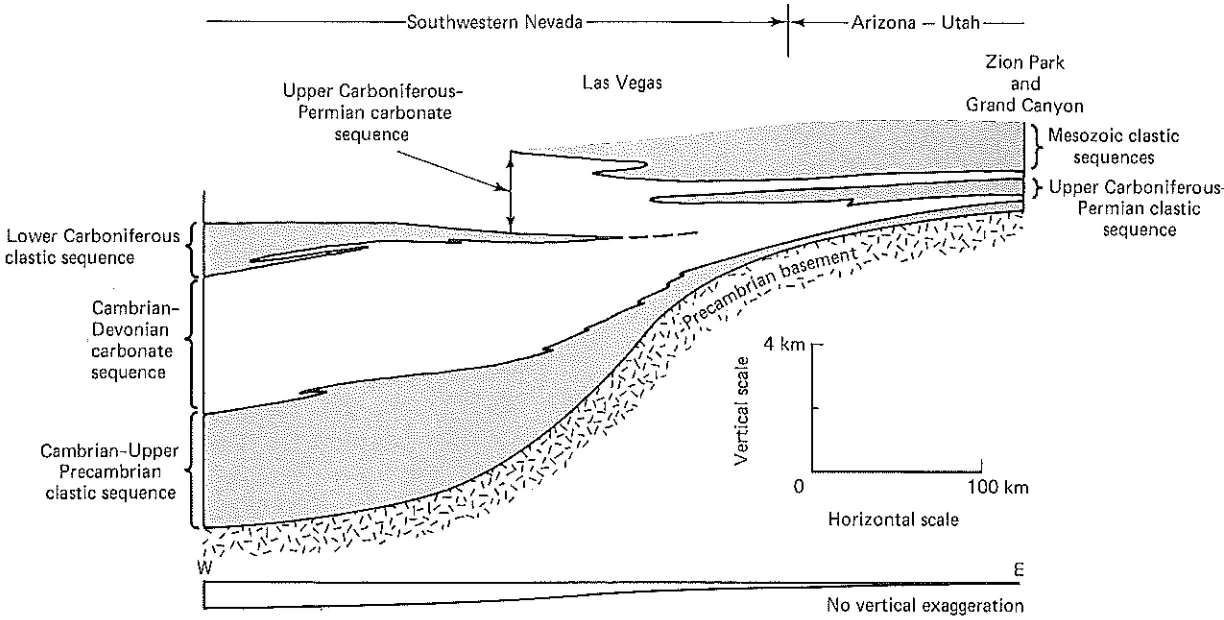

FIGURE 13–4 Restored stratigraphic section of eastern margin of Cordilleran geosyncline from the Grand Canyon to southwestern Nevada. See Figure 13-9 for location. (Simplified from Armstrong, 1968.)

Atlantic Ocean. The upper part of the Cordilleran Upper Precambrian clastic section and the overlying carbonate is fairly similar from the Yukon to California and even to the Appalachians, reflecting a stable continental margin whose sedimentation is dominated by changes in sea level and latitude. However, as we go deeper the section becomes much more heterogeneous, displaying rapid facies changes and evidence for syndepositional normal faulting and basaltic volcanism. The beginning of this episode of deposition is roughly 750 to 800 m.y. ago and is widely interpreted tectonically as a rifting event that established the late Precambrian and early Paleozoic stable continental margin of the Cordillera. It is interesting to note that the rifting of the Appalachian margin was at least roughly contemporaneous.

Only a fragmentary record of these earliest deposits of the Cordilleran stable continental margin and its Precambrian basement is exposed; however, the fragments are widely scattered from the Yukon to California. One of the places they are exposed is in the Cordilleran fold-and-thrust belt in Utah near Salt Lake City (Fig. 13-5). There in the Wasatch Mountains is a complete section extending from old Precambrian basement to Cretaceous at a paleogeographic position midway between the craton and the true Cordilleran geosyncline. The exposures

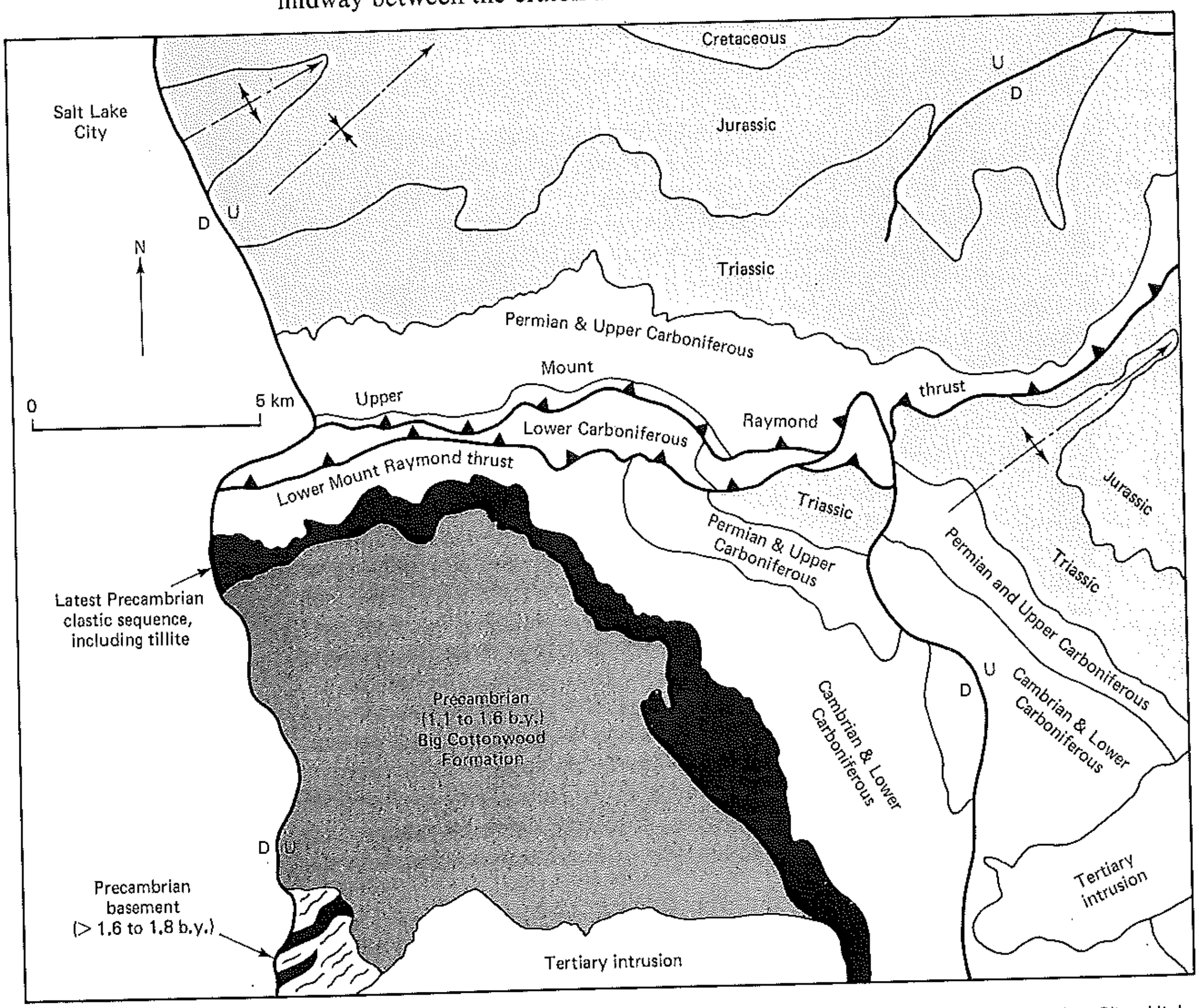

FIGURE 13–5 Simplified geologic map of Cottonwood area near Salt Lake City, Utah. (Compiled from mapping by M. Crittenden, U.S. Geological Survey.)

in map view are very unusual for the fold-and-thrust belt because the northeast-southwest-striking Cretaceous overthrust belt has been refolded about an east-west axis in the early Cenozoic Laramide orogeny (Fig. 13-24); as a result we see an oblique cross section of the overthrust belt in map view (Fig. 13-5). For example, the late Cretaceous Mount Raymond thrust has a slip vector that is toward the southeast, giving it a large component of strike slip in the present map view because of the refolding. Notice that it rides along a décollement in the Lower Carboniferous and then steps up eastward to the Jurassic, with an apparent offset of a few kilometers.

Our main concern for the moment, however, is the stratigraphy exposed in this oblique cross section because it gives a very deep look at the Cordilleran geosyncline. The deepest rocks exposed are gneissic and metamorphic Precambrian basement rocks, not less than 1.6 to 1.8 b.y. old. They are some of the few exposures of old Precambrian crust of North America within the Cordilleran mountain belt. Nevertheless, this old crust is known to extend far to the west in the Cordillera based on a remarkable isotopic property of Mesozoic and Cenozoic igneous rocks of the Cordillera. Magmas that pass through old Precambrian crust on their way up from the mantle take on a trace of the distinctive isotopic composition of old continental crust, whereas magmas that are west of the edge of Precambrian continental crust show isotopic compositions much closer to mantle compositions. Strontium isotopes are the most-useful tracer because ^{87}Sr is produced by decay of ^{87}Rb and rubidium is in the crust; therefore, old crust has a lot of ^{87}Sr. Samples of igneous rock with high ratios of ^{87}Sr to ^{86}Sr are considered to reflect Precambrian crust at depth. The line of $^{87}Sr/^{86}Sr > 0.704$ shown in Figure 13-11 is often taken as the western limit of the old crust and, therefore, the edge of the continent in latest Precambrian and early Paleozoic time. This edge is far to the west of our outcrops of this crust in the Wasatch Mountains to which we now return (Fig. 13-5).

There is still more tectonic history in western North America that predates the Cordilleran rifted continental margin; this history is represented by the 4-km-thick Big Cottonwood Formation that overlies the old metamorphic basement. The Big Cottonwood Formation is just one of many thick Upper Precambrian (1.1 to 1.8 b.y.) clastic formations that are exposed along the western margin of the continent in Canada and the United States. The most widely used names are Belt and Purcell Supergroups. These formations fill grabens and more extensive rift systems and for this reason have been thought by some people to be the initial rift deposits of the Cordilleran continental margin. It is more likely, however, that they record an earlier period of extension. The main evidence is that the rifting and thermal subsidence of a stable continental margin takes about 200 m.y. (Chapter 1); therefore, the Big Cottonwood Formation and its equivalents are 0.5 to 1.2 b.y. too old to be the initial deposits of the latest Precambrian–early Paleozoic Cordilleran continental margin. This observation shows us that the continental crust that preceded the Cordilleran mountain belt was a collage of very thick, sediment-filled grabens and older metamorphic and igneous basement.

As we pass upward through the latest Precambrian clastic section, we find marine and terrestrial glacial deposits, which are widespread on the edges of North America and some other continents at this time. The deposits in the Wasatch Mountains are largely nonmarine and the unconformities on the map reflect the filling of late Precambrian glacial valleys. After the glacial periods, there was the Cambrian transgression, followed by a long period of stable deposition. In this area there is no obvious sign of tectonism between the time of the late Precambrian glaciation and the Cretaceous thrusting, a period of a half a billion years. We see a conformable stratigraphic sequence on the map (Fig. 13-5), exhibiting disconformities reflecting only sea-level and epeirogenic effects. It is deceptive, however, to try to view the history of an orogenic belt from a single geographic position. In fact, deformation was going on from time to time not far away, beginning as early as late Devonian and early Carboniferous.

THE ANTLER OROGENY (Late Devonian and Early Carboniferous)

The first sign of major Phanerozoic orogeny in the Cordillera is given by a wedge of shale, sandstone, and conglomerate derived from the west in latest Devonian and early Carboniferous (Mississippian) time. This clastic wedge represents a marginal effect of the *Antler orogeny* and can be seen west of Las Vegas in the stratigraphic diagram of Figure 13-4. The Antler clastic wedge is possibly present as far north as the Yukon territory in Canada, but it is best developed in central Nevada (Fig. 13-6), where the deformational cause of the detritus is also well exposed without an undecipherable amount of later deformation and metamorphism.

Central Nevada is part of the Basin-and-Range province marked by major late Tertiary to Recent normal faulting; nevertheless, fragments of the Paleozoic

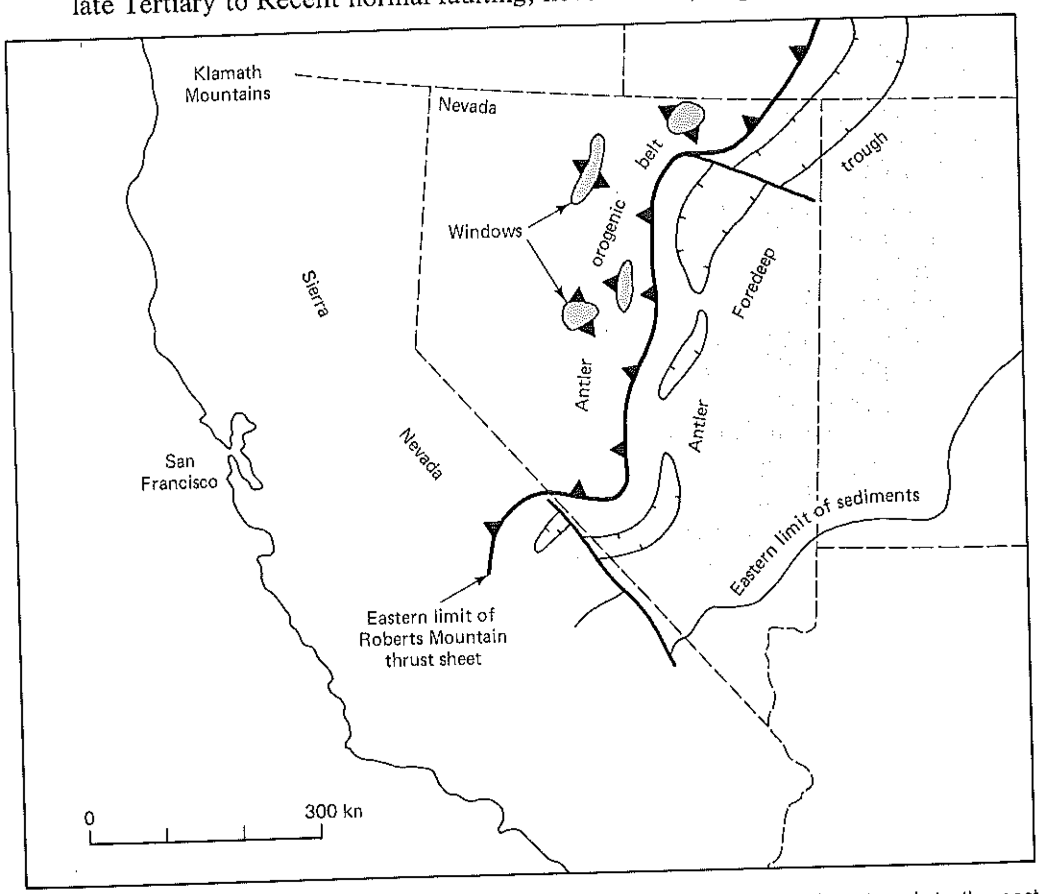

FIGURE 13–6 The Antler orogenic belt in central Nevada and foredeep trough to the east showing the extent of upper Lower Carboniferous strata. Windows through the Roberts Mountain thrust sheet expose autochthonous lower Paleozoic shallow-water carbonates, whereas the structurally overlying rocks are lower Paleozoic deep-water sediments. (Data from Stewart, 1980, and Poole and Sanberg, 1977.)

history are displayed in the mountain ranges, which are the horsts. We can trace the Cambrian through Devonian carbonate-bank sequence much of the way across Nevada, going from horst to horst. In the east the carbonate section is overlain by the westward-thickening and coarsening Antler clastic wedge, but as we move farther west we begin to find mountain ranges displaying thrust sheets of complexly deformed slate, chert, and argillite that is structurally overlying the carbonate sequence. A cross section of one of these ranges is shown in Figure 13-7. The carbonate sequence is shown at depth, exposed at the surface in a window on the west side of the range, with an overlying complex of coeval Ordovician through Devonian deep-water clastic sediments deformed in a number of east-vergent thrusts, many of which are refolded. Later igneous intrusions and normal faults of Tertiary age post-date the thrusting. The basal thrust of the allochthonous deep-water sequence is usually called the *Roberts Mountain thrust* and has been traced through many of the ranges of central Nevada. Stratigraphic relations show that it was emplaced during the time of the Lower Carboniferous (Mississipian) clastic wedge because the thrust sheet contains deformed Devonian strata and is unconformably overlain by Upper Carboniferous (Pennsylvanian) and Permian. The Roberts Mountain thrust sheet is an important structural manifestation of the Antler orogeny.

What is the larger tectonic significance of this late Devonian and early Carboniferous Antler orogeny? There is so much later tectonic disruption of the Cordillera that complete understanding may not be possible, but several conclusions can be made. First, sedimentological and stratigraphic studies show that there is an interfingering facies relationship between the autochthonous carbonate-bank sequence and the deep-water allochthonous clastic sequence (Silberling and Roberts, 1962). Apparently the Roberts Mountain allochthon is a deep-water slope-and-rise sequence, in many ways analogous to the Taconic sequence of the Appalachian mountain belt. We can conclude that some compressive plate boundary impinged on the stable continental margin of the western United States in late Devonian and early Carboniferous time and caused the thrusting of the continental rise-and-slope sediments onto the carbonate shelf.

An exact plate-tectonic scheme for the Antler orogeny is speculative. Perhaps the Antler orogeny is the result of arc-continent collision, by analogy with the Taconic orogeny of the Appalachians. This is a likely possibility and a popular interpretation, but the necessary supporting data are less complete than for the Taconic orogeny. Possible candidates for the island arc and other associated oceanic rocks are present in the Sierra Nevada and Klamath Mountains (Fig. 13-6) of California to the west (Burchfiel and Davis, 1975; Schweickert and Snyder, 1981). The principal uncertainties stem from the fact that many later orogenic events are known in this region, some of which are discussed later in this chapter; these orogenies show signs of producing such profound displacements that the oceanic Paleozoic rocks of California may have nothing to do with the Roberts Mountain thrust sheet and the Antler orogeny. We shall not pursue this uncertain path; many such paths exist in this and all mountain belts. Our concern in this chapter is the better-established aspects of the Cordilleran system. What is clear about the Antler orogeny is that it represents the destruction of the late Precambrian through Devonian stable continental margin and marks the beginning of Cordilleran orogenesis, at least in the western United States.

The Antler orogeny was not long lived, lasting less than 25 m.y. in late Devonian and early Carboniferous time. By late Carboniferous time, the orogenic belt appears to have been well eroded and began to be covered unconformably by a new cycle of stable continental margin sedimentation.

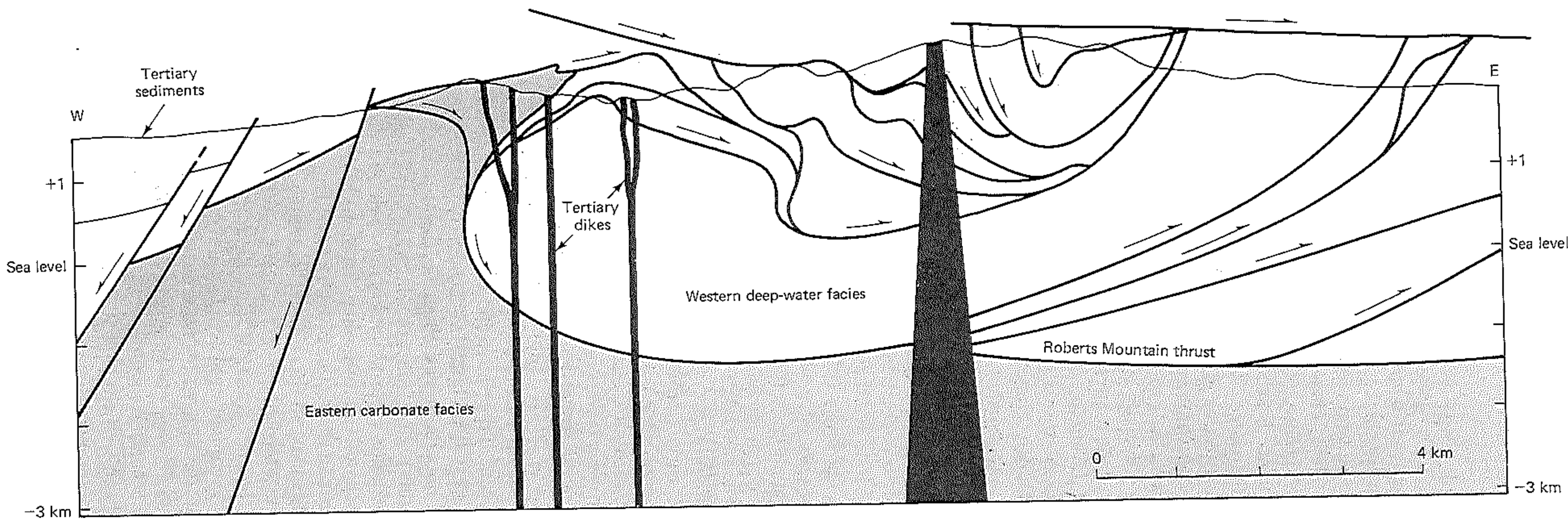

FIGURE 13–7 Cross section of early Carboniferous Roberts Mountain thrust sheet, Shoshone Range, Nevada. Thrust imbrications display folding during thrusting. Tertiary structure includes normal faulting and igneous intrusion. (Simplified after Gilluly and Gates, 1965.)

ANCESTRAL ROCKY MOUNTAINS (Late Carboniferous and Permian)

The part of the Cordilleran mountain belt exposed in the western United States is remarkable in that it displays a fairly clear record of a number of discrete orogenic events, extending as far back as the Antler orogeny. They are spread out spacially to a sufficient extent that younger deformation, metamorphism, and magmatism does not extensively obscure older events. In contrast, Tertiary magmatism and Mesozoic deformation, sedimentation, and magmatism have been so extensive in the Mexican Cordillera that Paleozoic and Precambrian rocks are exposed only in small enclaves and record only a very fragmentary pre-Mesozoic history. The late Paleozoic deformational history of the Canadian Cordillera is also fairly obscure for reasons we shall see later. Therefore, we continue our discussion in the U.S. Cordillera but, perhaps surprisingly, move nearly a thousand kilometers east, well into the Paleozoic craton near Denver, Colorado, to find the next locus of Paleozoic orogenic activity.

The continental interior east of the Cordilleran miogeosyncline had been an area of cratonic stability, exemplified by the Grand Canyon (Fig. 13-3), since the Cambrian. During the Antler orogeny in the early Carboniferous, the continental interior was a stable cratonic carbonate sea unaffected by the deformation to the west in Nevada; to the southeast in the Ouachita orogenic belt of Arkansas, Oklahoma, and the Gulf Coast; or to the north in the Innuitian orogenic belt of the Canadian Arctic.

Seemingly without warning, the locus of deformation stepped out into the continental interior in late Carboniferous (Pennsylvanian) time with the deformation of the Precambrian basement in a system of fault blocks, largely marked by major strike-slip faults with associated thrust and normal faults. We see the effects of this deformation in the Rocky Mountain area of Colorado as a major unconformity in areas of fault-block uplift and as thick orogenic sedimentation in the adjacent basins. Only in a few local areas have the actual late Paleozoic faults been clearly identified because the fault blocks partly coincide with the basement uplifts that form the present-day Rocky Mountains and with the latest Cretaceous–earliest Tertiary block uplifts of the Laramide orogeny. The late Carboniferous structures are called the *Ancestral Rocky Mountains* because of this partial coincidence with the present-day mountains.

What is known of the regional pattern of Ancestral Rocky Mountain deformation is summarized in Figure 13-8. Two main belts of block uplift are known, based on associated coarse clastic rocks fringing them in the adjacent basins and based on unconformities with overlying Permian or Triassic sediments. The eastern belt of uplifts includes the Ancestral Front Range, Apishapa, and Sierra Grande uplifts, which in part coincide with the present-day Front Range west of Denver but also are buried under the Great Plains of southeastern Colorado and northeastern New Mexico. The Apishapa uplift and the Sierra Grande uplift to the south are known largely from subsurface exploration. The coarse conglomeratic Upper Carboniferous sediments that flank the east side of the Ancestral Front Range uplift are exposed as the steeply dipping red beds of the well-known Fountain Formation in the western suburbs of Denver on the flanks of the present Front Range.

The second major uplift is the Ancestral Uncompahgre uplift, which in part coincides with the present-day Uncompahgre Plateau; most of the late Paleozoic uplift is delineated by subsurface exploration. Its age appears to be slightly younger than the Ancestral Front Range based on the sedimentary history of the flanking basins, apparently latest Carboniferous and Permian as opposed to middle and late Carboniferous. The intervening Colorado trough is a site of

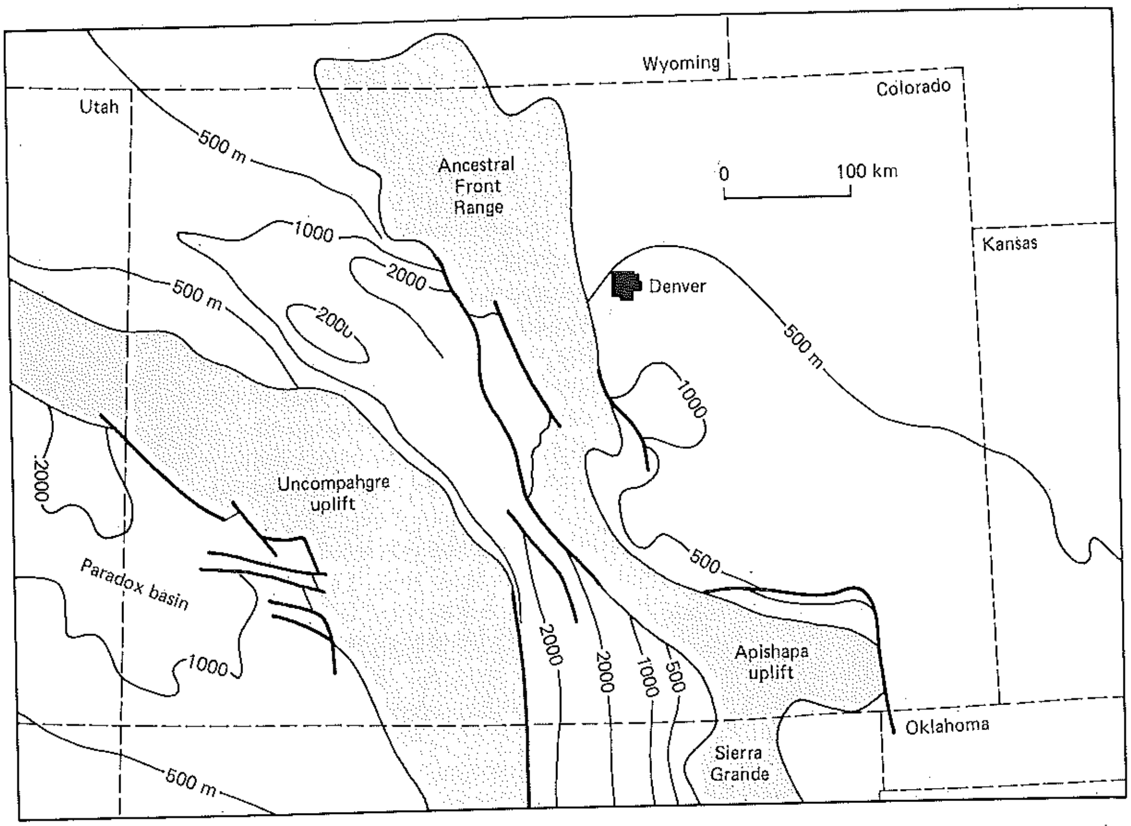

FIGURE 13–8 Uplifts, adjacent basins, and faults of the late Carboniferous and early Permian Ancestral Rocky Mountains in Colorado. Thickness of Upper Carboniferous strata in meters. (Simplified after De Voto, 1980.)

massive orogenic sedimentation amounting to more than 3000 m in the late Carboniferous and Permian. The sediments of the Colorado trough are now exposed in several of the present Rocky Mountain uplifts. By Permian time the orogenic source of sediment had diminished substantially, so that the orogenic troughs began to receive finer sediments, including substantial evaporites in the Paradox basin, southwest of the Uncompahgre uplift (Figs. 13-8 and 7-38).

The Ancestral Rocky Mountains are not a wholly Cordilleran orogenic event. This zone of basement deformation can be traced eastward through Oklahoma and Texas by drilling and seismic methods as a series of buried mountain ridges and flanking basins, which are important for their petroleum deposits. Some of them finally come to the surface in the low Arbuckle and Wichita Mountains of southeastern Oklahoma, where this zone of late Carboniferous and Permian basement deformation impinges on the Ouachita orogenic belt north of Dallas (Fig. 12-1). The Ancestral Rocky Mountains are, therefore, part of a zone of deformation that cuts across the North American craton between the Ouachita and Cordilleran geosynclinal mountain belts.

Other zones of deformation of the craton are present at the same time elsewhere in North America, including the Kentucky–Rough Creek fault zone extending west from the Appalachians into the Illinois basin, the Nemaha ridge in eastern Kansas and Nebraska, and an early Devonian zone of basement deformation extending from the Innuitian mountain belt of the Canadian Arctic Islands south across the craton to the center of Hudson's Bay. Apparently North America was caught between active mountain belts on so many sides that the cratonic interior began to fail, much like central Asia today north and east of Tibet, which is cut by zones of active basement block uplift, marked by major strike-slip, with associated thrust faults with flanking deep basins and grabens that are forming in response to the collision of India and Eurasia. The Ancestral Rocky Mountains and the Arbuckle and Wichita mountains are apparently an analogous effect of collision between South America and North America along the Ouachita orogenic belt.

SONOMA OROGENY (Late Permian and Early Triassic)

The deformation of the normally stable continental interior represented by the Ancestral Rocky Mountains is somewhat exceptional, but by no means unique; this region was again subjected to major deformation in the latest Cretaceous and early Tertiary Laramide orogeny. Nevertheless, the continental margin is the normal site of orogenic instability; therefore, it is not surprising that we must retreat from the unstable craton to the now-familiar Basin and Range of central Nevada to observe the next major orogenic event, the late Permian and early Triassic *Sonoma orogeny*.

During the Sonoma orogeny, deep-water upper Paleozoic sedimentary and volcanic rocks of the *Golconda allochthon* were thrust from the west over a coeval autochthonous shallow-water sequence. The thrust contact is exposed in many of the horsts of central Nevada (Fig. 13-9) and is commonly called the *Golconda thrust*. Stratigraphic relations exposed in some of the ranges show that the time of emplacement of the Golconda allochthon must be within the late Permian or early Triassic because deformed sediments of both the autochthon and allochthon include widespread Permian strata and the Golconda thrust is overlain unconformably by Lower and Middle Triassic sediments.

The autochthonous upper Paleozoic shallow-water sequence is very thin, generally only a few hundred meters thick, and characterized by chert-pebble conglomerates, sandstones, and limestones. This sequence is called the *Antler overlap sequence* because it unconformably overlies the deformed rocks of the earlier Antler orogeny, the chert-pebble conglomerates being derived from the oceanic rocks of the Roberts Mountain allochthon. The stratigraphy of the overlap sequence suggests that in late Carboniferous and Permian time, between the Antler and Sonoma orogenies, the continental margin ceased to be a site of active deformation, for a period of about 100 m.y. Therefore, the Sonoma orogeny marks a major new episode in the progression of the Cordilleran mountain belt. In essence, what happened?

The next important observation is that the Golconda allochthon includes continental-slope-and-rise sediments that can be correlated with the autochthonous-overlap sequence. Furthermore, the other important rock units of the Golconda allochthon are pelagic oceanic rocks and widespread volcanic rocks of island-arc affinities, including pyroxene andesite, basalt, and dacite, with associated volcanogenic sediments and shallow-water limestones. Therefore, the rocks deformed and thrust onto the continental shelf in the Sonoma orogeny include a stable continental-margin sequence, overlapping the roots of the Antler orogenic belt, together with oceanic sediments and an island arc. It is generally concluded, based on these observations, that the plate-tectonic impetus for the Sonoma orogeny was an arc-continent collision. The Sonoma, Antler, and Taconic orogenies are broadly analogous in that each involves deformation of a previously stable continental margin by the thrusting of deep-water slope-and-rise sediments over the continental shelf, apparently in response to arc-continent collision.

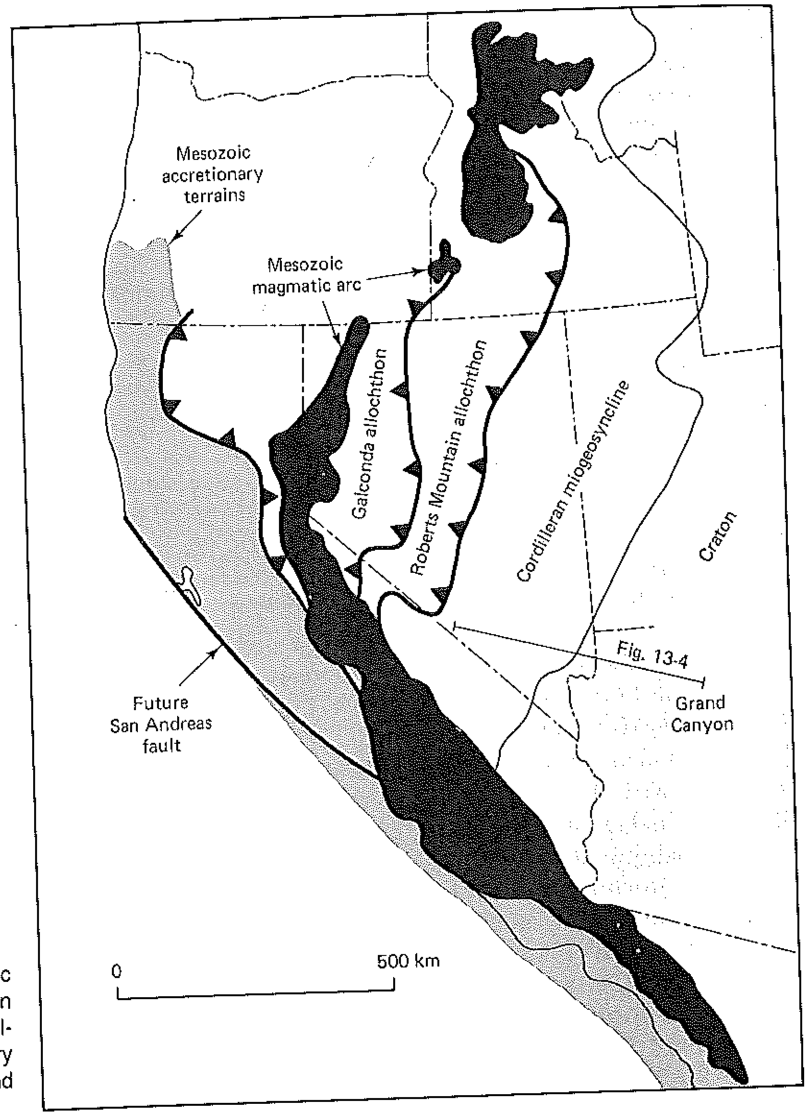

FIGURE 13–9 Truncation of Paleozoic structural trends in the southwestern United States by Mesozoic continental-margin magmatic arc and accretionary wedges. (Compiled from Burchfiel and Davis, 1975.)

EARLY MESOZOIC TRUNCATION EVENT

The Paleozoic continental margin in the western United States has a northeast-southwest orientation. For example, the boundary between the thin cratonic sequences such as the Grand Canyon and the thick cordilleran miogeosyncline, shown in cross section in Figure 13-4, has a N40°E strike (Fig. 13-9). Isopach maps showing the thickness variation of individual Paleozoic stratigraphic units show this same northeast-southwest orientation of the contours, reflecting the orientation of the Paleozoic continental margin (Stewart, 1980). The late Paleozoic continental-margin Antler and Sonoma orogenies also show this same orientation for the foredeep sedimentary trough and for the fronts of the Roberts Mountain and Golconda allochthons (Figs. 13-6 and 13-9). However, by late Jurassic, a new northwest-striking Mesozoic and Cenozoic structural grain, which cuts across and truncates the northeast-southwest Paleozoic structural grain, was well established. This structural reorganization of the western United States sometimes is called the *early Mesozoic truncation event*.

The truncation event not only represents a reorientation of continental-margin deformation, but it also marks the beginning of a new era of Cordilleran tectonics that is dominated by continental-margin volcanic and plutonic activity. Some of this magmatic activity is illustrated in Figure 13-9, which shows the belt of Mesozoic and early Cenozoic batholiths that exists along the western Cordillera. It is this belt of continental-margin magmatism and related deformation and sedimentation that cuts across the Paleozoic structural trends. In northern California the batholiths are emplaced into Paleozoic oceanic rocks related to the Antler and Sonoma orogenies. In central California the batholiths are emplaced into the Cordilleran miogeosyncline. In southern California and Arizona the batholiths are emplaced into the Paleozoic craton. The Mesozoic batholithic belt therefore cuts across the Paleozoic fabric of the western United States.

A second important observation is that the Paleozoic belts do not reappear on the southwest side of the batholith belt. Instead, we find belts of deformed Mesozoic continental-margin and oceanic rocks oriented northwest-southeast, parallel to the batholith belt (Fig. 13-9). Therefore, the continuation of the Paleozoic continental margin is apparently missing. Many workers have interpreted these observations to mean that the southwestward continuation of the Paleozoic continental margin was actually broken off and drifted away sometime in the Triassic or Jurassic. This is why the early Mesozoic tectonic change is called a *truncation event.*

This hypothesis of tectonic truncation is actually rather difficult to evaluate fully. The largest difficulty stems from the lack of extensive exposures of Paleozoic sediments and Precambrian basement rocks in the Mexican Cordillera. If the Paleozoic continental margin was not truncated in the early Mesozoic, it must have originally taken a sharp bend to the southeast in southern California and northwestern Mexico. This constitutes a second hypothesis, which is not as popular, partly in light of the available data on the few scattered exposures in Mexico but also based on considerations of the early Mesozoic plate tectonics of the North Atlantic and Gulf of Mexico (Anderson and Schmidt, 1983), which are outlined in the following paragraphs. By taking this larger view, we gain insight into the two aspects of the truncation event: (1) the actual truncation and (2) the beginning of Mesozoic continental-margin arc magmatism. Furthermore, we gain some insight into the role of plate tectonics in theories of ancient mountain belts.

By the end of the Paleozoic, North America, Europe, Africa, and South America were sutured together into one large continent along the Appalachian-Ouachita mountain belt. In late Triassic and early Jurassic time, parts of that suture began to open up again: the North Atlantic between Africa and North America and the Gulf of Mexico between North and South America. It should be noted that the North Atlantic between Europe and North America, the South Atlantic between Africa and South America, and the Caribbean are all later ocean basins. The major late Triassic and Jurassic rifting event is a splitting of North America–Europe from South America–Africa. It is generally considered significant that this time of major tectonic change between late Permian and Jurassic on the southeastern and eastern edge of North America is also a time of major tectonic change in the western Cordillera (Coney, 1978). Apparently, late Permian to early Jurassic was a time of significant absolute change in North American plate motion that affected plate interactions on both sides of the continent. This change in plate motion may account for the beginning of continental-margin arc magmatism in the Cordillera in late Triassic and Jurassic time.

The actual truncation of Paleozoic structural trends is also hypothesized by many to be a manifestation of the early Jurassic rifting. The reasoning is as follows. When we carefully attempt to fit the continents back together across the North and South Atlantic, we always find that there is an unsatisfactory overlap between South America and North America. Mexico and Central America are too far south for northern South America to fit neatly into the Gulf of Mexico, where it

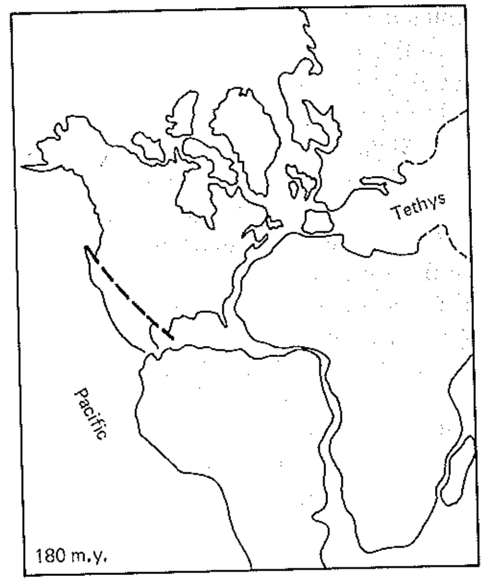

FIGURE 13–10 Reconstruction of early Jurassic positions of Eurasia and Pangea, showing problem of overlap of North and South America.

apparently belongs (Fig. 13-10). For this reason the Paleozoic and Precambrian continental crust of Mexico and Central America is generally thought to have moved southeast relative to the rest of North America as several small crustal blocks. The truncation of Paleozoic structural trends (Fig. 13-9) is hypothesized to be a result of this southeastward motion of Mexico; the bounding fault is proposed to be a hidden strike-slip fault system roughly along the U.S.–Mexican border extending between southern California and the Gulf of Mexico (Anderson and Schmidt, 1983).

In this discussion of the early Mesozoic truncation event, we have gone far beyond the basic geologic observations of the western Cordillera in search of some explanation of what happened between the Late Permian and Late Jurassic. What is clear is that some major change in tectonic orientation and style took place (Fig. 13-9). It is less clear why. Researchers in regional structural geology of mountain belts generally look to plate tectonics as the most-important cause of the major tectonic changes that are observed and documented in mountain belts. The plate-tectonic explanation of the early Mesozoic truncation event is a good example of the sorts of attempts at explanation that are made; the attempt was successful because it was able to join together otherwise unrelated observations into a comprehensive picture, but it is uncertain because all the necessary geologic data are not available. Attempts at plate-tectonic explanation generally become more satisfactory as the mountain belt becomes younger. For example, in our brief introduction to the present-day spatial changes in tectonic style along the Cordillera (Fig. 13-1), most of the important present-day tectonic changes within the mountain belt could be easily related to changes in plate boundaries along the continental margin. In contrast, map-view plate tectonic explanations of the Paleozoic Appalachian mountain belt are so uncertain that we did not present any in Chapter 12. As we proceed to the late Mesozoic and Cenozoic tectonics of the Cordillera, plate-tectonic explanations in general become more specific, testable, and enlightening.

PACIFIC-MARGIN MAGMATIC BELT

The Pacific margin of the Cordillera, particularly in the United States, was an active site of continental-margin subduction quasi-continuously from late Triassic until late Cenozoic. This fact is most easily documented by study of the volcanic and plutonic rocks of the Pacific margin, including the batholith belt (Fig. 13-9). Three types of evidence are particularly important.

1. Petrographic and geochemical studies of the magmatic belt show that it has close petrologic affinities to present-day continental-margin volcanic arcs such as the Andes, Cascades, and Indonesia.
2. Field studies show that many of the magmatic rocks were intruded into the North American continental margin or were erupted upon it. Other magmatic rocks—for example, in the Jurassic of the western Sierra Nevada—may represent oceanic island arcs that are accreted to North America; however, there also are coeval arc rocks to the east that were clearly emplaced on the North American continent (Schweikert, 1981; Burchfiel and Davis, 1981). The emplacement of these Cordilleran granitic plutons has already been discussed in Chapter 7 (Figs. 7-20 and 7-23).
3. Isotopic and fossil dating show that the continental-margin magmatic arc was active for much of the time between late Triassic and late Cenozoic. However, the history has not been uniform and monotonous. The available data show that there are several episodes of especially abundant and widespread magmatism, whereas other times show little activity. Subduction was certainly not uniform and may not have been continuous. The data also show that the locus of magmatic activity has undergone important changes in map pattern during the last 200 m.y. Several of these changes in abundance and location of the Cordilleran magmatic arc are illustrated in the following paragraphs.

Two periods of late Mesozoic magmatism were especially important in constructing the batholith belt in the western United States and northwestern Mexico. This fact can be illustrated by histograms of potassium-argon dates on plutonic rocks of the western United States (Fig. 13-11). So many plutonic rocks have been isotopically dated in the western United States that they give a reasonably accurate impression of the temporal pattern of late Jurassic through early Cenozoic magmatic activity. The pre-late Jurassic history is largely obscured because K-Ar dates only give the time of last cooling below about 300° to 400°C, and many of the older rocks have suffered reheating after their initial crystallization. Figure 13-11 strictly gives a picture of times of major cooling. The mid-to-late Cenozoic igneous history also is not well represented on the plutonic histograms because most of this younger history is represented at the surface by volcanic rocks, which is discussed later. Two periods of especially voluminous magmatic activity are immediately apparent in Figure 13-11: late Jurassic and early Cretaceous (about 130 to 160 m.y.) and late Cretaceous (about 75 to 105 m.y.).

The late Jurassic and earliest Cretaceous magmatic activity is one manifestation of a period of widespread tectonic activity sometimes called the *Nevadan orogeny*. The late Cretaceous period of tectonic activity is sometimes called the *Sevier orogeny*, although this name is less universally applied. The period of latest Cretaceous and earliest Cenozoic magmatism that appears in southern California, Arizona, and the Pacific Northwest, but not in the Sierra Nevada, is a manifestation of the *Laramide orogeny*. Later in this chapter we shall see that each of these three periods of orogenic activity display important tectonic phenomena in other geologic provinces of the Cordilleran mountain belt in addition to the magmatic arc.

The names for each of these three orogenic periods were originally applied to locally recognized orogenic phenomena, just as *Taconic orogeny* was originally

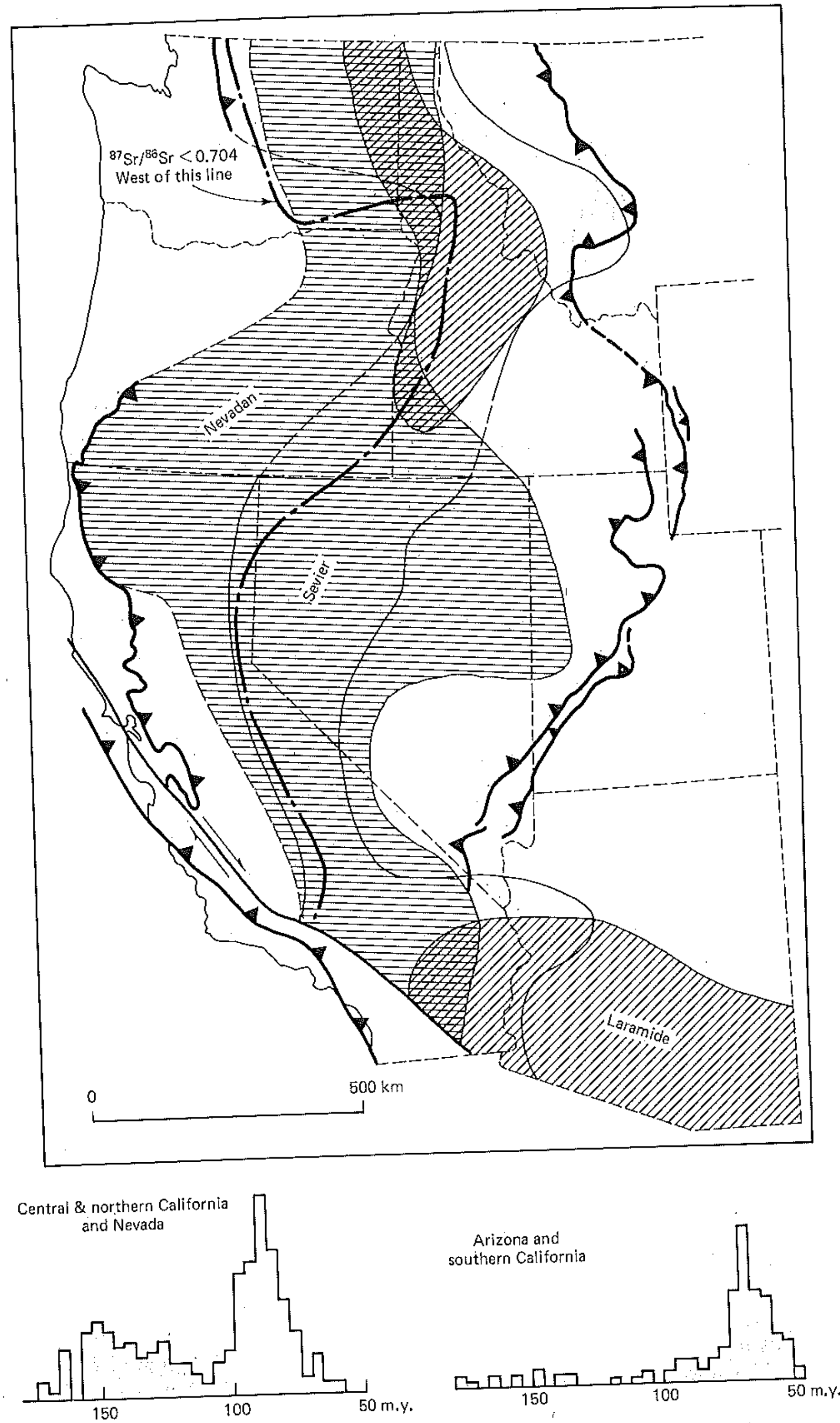

FIGURE 13–11 Loci of late Jurassic to early Cretaceous Nevadan (160 to 125 m.y.) mid-to-late Cretaceous Sevier (105 to 75 m.y.) and Laramide (50 to 75 m.y.) magmatism in the western United States. The histograms give K/Ar radiometric dates on the granitic rocks. The presumed western edge of Precambrian Continental crust is along the line defined by $^{87}Sr/^{86}Sr = 0.704$ for Cenozoic volcanic rocks and Mesozoic plutonic rocks. (Based on data of Armstrong and Suppe, 1973; Kistler and Peterman, 1973; and Armstrong, Taubeneck, and Hales, 1977.)

applied to an angular unconformity along the Hudson River in New York State (Fig. 12-10). *Nevadan orogeny* was originally applied to an angular unconformity in the Sierra Nevada and Klamath Mountains of California. *Sevier orogeny* was originally applied to the fold-and-thrust belt of western Utah. *Laramide orogeny* was originally applied to the basement-block uplift of the Laramie Range of central Wyoming and other similar nearby uplifts. In this chapter we apply these orogenic names in their broader senses to help us identify some of the major throughgoing features of Cordilleran tectonics.

The spatial distribution of magmatic activity within the arc is quite different during each of the three orogenic periods mentioned above and is illustrated in Figure 13-11. During the late Jurassic, Nevadan orogeny magmatic activity was very widely distributed across the western United States from the Pacific margin in California to the edge of the Paleozoic craton in Utah. During the late Cretaceous Sevier orogenic period, magmatism was more nearly confined to the 150-km-wide belt of closely spaced intrusions of the Sierra Nevada and Peninsular Range batholiths of California and Baja California. Much of the volume of these large composite batholiths was emplaced during a relatively short time in the late Cretaceous, creating a vast sea of granitic plutonic rock. The central zone of the batholith belt is nearly all late Cretaceous granitic rock. For example, most of the area of the central Sierra Nevada shown in Figure 13-12 is late Cretaceous granitic rock. Just before the end of the Cretaceous, there was a major change in the map distribution of magmatism (Fig. 13-11); this change marks the beginning of the Laramide orogenic period.

The early Jurassic and younger arc magmatism on the Pacific margin of North America has many petrologic similarities to the present-day continental-margin magmatism of the Andes, which points to subduction that is localized along the edge of the continent. For this reason the Cordilleran continental margin after the early Mesozoic truncation event has often been called an *Andean-type margin*. In contrast with the Andes, however, the Cordillera does not appear to have generally had the very high topography that is characteristic of much of the present-day Andes. During Jurassic time, volcanic rocks of the Cordilleran arc generally interfinger with shallow-water marine sediments. Some of the volcanic centers were actually submarine calderas. Possible analogs of this continental-margin arc magmatism at or near sea level include the present-day arcs of western Indonesia and Central America south of Guatemala. The magmatic arc also appears to have been at low elevation during most of the Cenozoic in the western United States based on evidence from fossil floras interbedded with the volcanic rocks. Local mountains, especially volcanoes, may have reached substantial elevations, but the regional topography was low. The present high elevation of the western United States is a late Cenozoic orogenic phenomenon, to be discussed near the end of this chapter. The only period of possible Andean topography to the western United States is the late Cretaceous, and even that is unlikely based on present data.

PAIRED CRETACEOUS BASEMENT BELTS

In the brief introduction to Cordilleran tectonics presented in this chapter, we skipped over essentially all the late Triassic and Jurassic tectonism of the Pacific margin, including much of the Nevadan orogeny. This period is represented in the rock record by some extremely complex and fragmented geology that is the focus of very active research (for example, see Ernst, 1981). Interpretations of the tectonics of this period are controversial and insight is rapidly evolving. We summarize this period by simply noting that a number of belts of oceanic rocks,

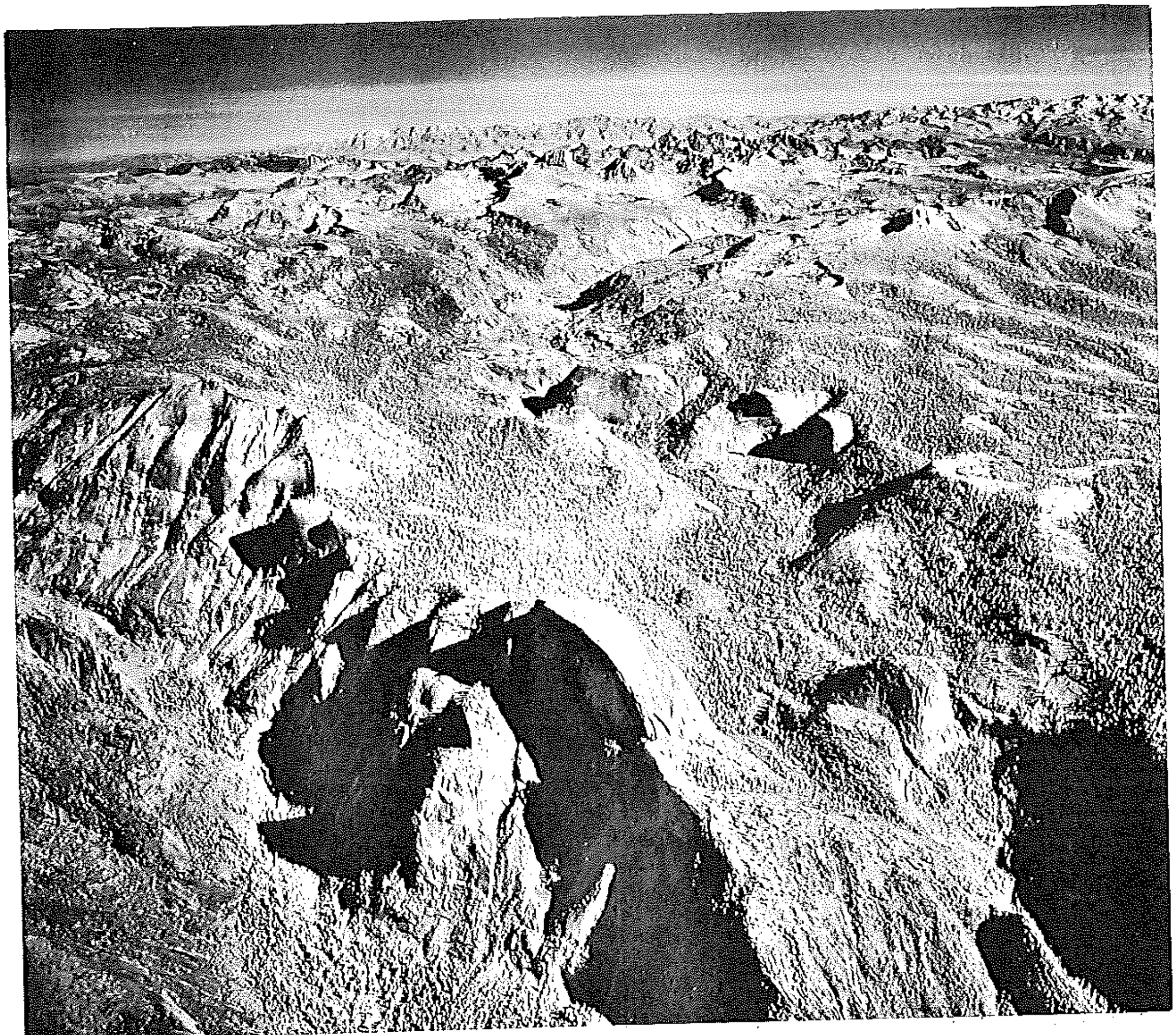

FIGURE 13–12 Oblique aerial view of Sierra Nevada batholith belt, Yosemite National Park, looking east. All the foreground and middle ground is underlain by Cretaceous granitic rock. The darker peaks in the background are underlain by metamorphic roof pendants of the Ritter Range, which contains extrusive equivalents of the mesozoic plutons. The peak in the foreground is Half Dome (see also Fig. 6-21). (U.S. Geological Survey.)

including ophiolites, pelagic sediments, and island-arc volcanic rocks, were accreted to parts of western North America during this period. We now move on to the Cretaceous of Oregon, California, and northwestern Mexico, which represents a time of great unity and throughgoing tectonics in the Cordilleran margin.

The title of this section, "Paired Cretaceous Basement Belts," provides a hint of the fact that one of the great unifying features of Cretaceous Cordilleran tectonics is a pair of belts of contrasting basement rocks running along the North American continental margin; they extend for more than 3000 km from southern Baja California to Oregon. In this context we mean by *basement* any metamorphic and plutonic rocks that act as basement for the overlying unmetamorphosed Cenozoic sedimentary and volcanic rocks.

The Cretaceous paired basement belts (Fig. 13-13) comprise (1) the *batholith belt,* already discussed, with its high-temperature, low-pressure metamorphic

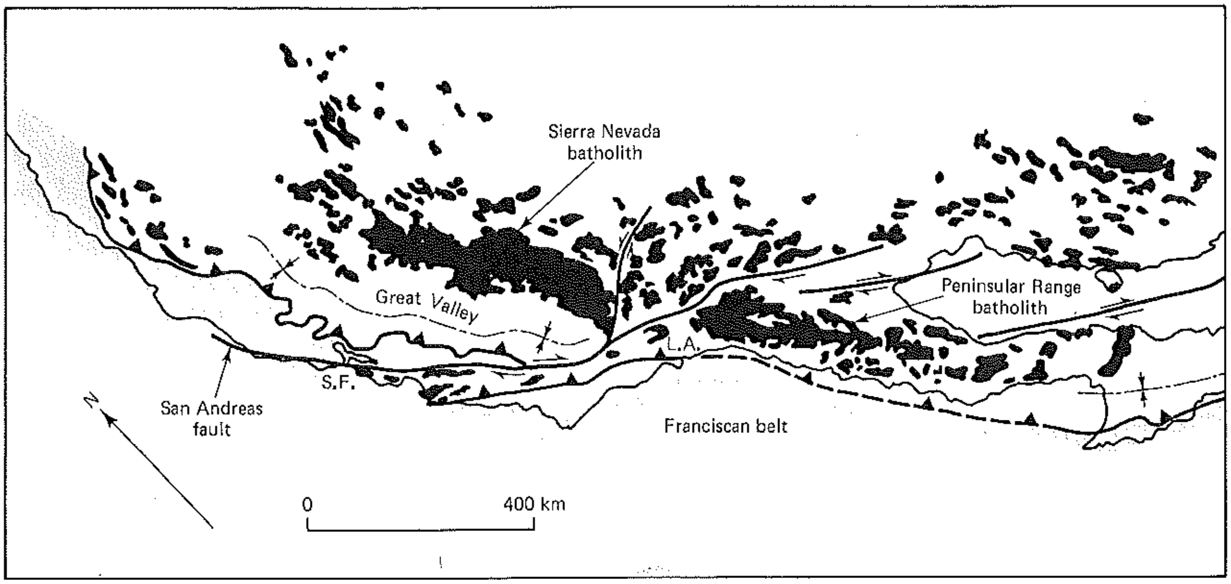

FIGURE 13–13 Cretaceous paired-basement belts, Oregon to Baja California. The high-pressure, low-temperature Franciscan belt is shown by the screened pattern along the continental margin. The granitic rocks of the high-temperature, low-pressure batholith belt are shown as black splotches.

country rock and (2) the *Franciscan belt* of low-temperature, high-pressure metamorphic rocks of oceanic origin, which exhibits little magmatic activity other than oceanic basaltic volcanism prior to metamorphism. These two contrasting basement belts are the same age and run approximately parallel to each other and to the continental margin from Oregon to southern Baja California (Fig. 13-13). The original parallel pattern has been disrupted in part by the large Cenozoic strike slip on the San Andreas fault system that causes a repetition of the basement belts in map view in central California. In going west from the Sierra Nevada in central California, we go from the batholith belt to the Franciscan belt, cross the San Andreas fault, go back into rocks of the batholith belt, and then into the Franciscan belt again near the coast. In spite of these complexities caused by later deformation, a single pair of belts existed in the late Mesozoic, with the Franciscan belt to the west.

This pair of basement belts is not unique. Similar paired-basement belts are fairly common, particularly in the young mountain systems that ring the Pacific margin (Miyashiro, 1961). For example, they exist in New Zealand, Sulawesi, Japan, Kamchatka, southern Alaska, and southern Chile. One essential feature of paired belts is that the metamorphic and plutonic activity of the two halves is of the same age, whereas the stratigraphic age of the sedimentary protoliths may differ. For example, the metamorphic rocks of the Cordilleran batholith belt range in stratigraphic age from Precambrian to Cretaceous, whereas the high-temperature metamorphism and magmatic activity is late Mesozoic. In contrast, the Franciscan belt has a stratigraphic age ranging from Jurassic through the Cretaceous and displays late Mesozoic metamorphism.

A second essential feature of the paired-basement belts is that one belt is characterized by metamorphism that records a low geothermal gradient, whereas the other records a high geothermal gradient with abundant associated arc magmatism. The low-temperature, high-pressure Franciscan belt typically displays zeolite, blueschist, high-pressure amphibolite, and eclogite facies metamorphism, whereas the high-temperature, low-pressure belt may display hornfels and other high-temperature facies.

The two coeval belts of contrasting thermal gradient running parallel to the continental margin must reflect something fundamental in the underlying tectonic processes. We can discover what this is by looking to present-day analogs of paired-basement belts. We have already noted that the batholith belt exhibits strong petrologic similarities to continental-margin magmatic arcs. Studies of present-day geothermal gradients in the environs of magmatic arcs show that they have the high thermal gradients and heat flow required for present-day high-temperature, low-pressure facies of metamorphism. Seaward of the magmatic arc in the region of the accretionary wedge (Fig. 1-15) is a region of low thermal gradient and low heat flow; this is thought to be a site of present-day low-temperature, high-pressure metamorphism. Numerical thermal models show that the high thermal gradient of magmatic arc is due largely to magmatic convection of heat and hot upper mantle, whereas the low thermal gradient of the accretionary wedge is due to the subduction process. Subduction places cooler rocks under warmer rocks, thereby causing a reversal or reduction of the geothermal gradient; little heat flows into the accretionary wedge. Therefore, a good physical basis exists for drawing an analogy between late Mesozoic paired-basement belts of the Cordillera and present-day continental-margin island arcs. The paired-basement belts record the fact that the late Mesozoic continental margin was a site of subduction.

We now turn to the Franciscan belt for a more-detailed look. What we see indeed looks like what we would expect of a large continental-margin accretionary wedge (Fig. 1-15); the Franciscan belt includes a structurally complex assemblage of deep-oceanic and continental-margin rock types. The deep-oceanic rocks are fragments of ophiolitic material and overlying pelagic sediments, including serpentinite, gabbro, basalt, radiolarian chert, submarine hot-spring manganese deposits, and pelagic limestone. At least one normal-sized submarine seamount is exposed in a deformed and metamorphosed state in the Franciscan terrain (Snow Mountain Volcanics in Fig. 13-14). The continental-margin sediments include thick sequences of arc-derived turbidites and interbedded olistostromes; they compose much of the volume of the Franciscan terrain. These sediments are widely interpreted as trench deposits, although other oceanic paleogeographic settings near the continent may also be represented. Little is known of the specific paleogeography of the Franciscan ocean basin. In fact, so much time is involved (late Jurassic through middle Tertiary, 155 to 30 m.y.) that an enormous area must be represented; for example, if the average rate of subduction were only 5 cm/y, over 6000 km of paleogeography would have been accreted into the Franciscan wedge.

Another fundamental fact contributes to the paleogeographic complexity of the Franciscan ocean basin, namely, that it never existed in its entirety at any single time. One of the most important observations about the Franciscan belt is that deposition, deformation, and high-pressure metamorphism are all broadly coeval throughout the belt. Some rocks were being deformed and metamorphosed at great depth at the same time that other, now nearby, rocks were far out to sea being deposited in an oceanic environment. For example, in the part of the northern Coast Ranges shown in Figure 13-14, some of the metamorphic rocks at Goat Mountain were undergoing high-pressure metamorphism for about 20 m.y. in the late Jurassic (155 through 135 m.y.) at the same time as the sedimentary and volcanic sequence on Snow Mountain (Fig. 13-14) was being deposited. At Snow Mountain we find a late Jurassic sequence of pelagic and abyssal-plain sediments overlain by an oceanic seamount volcano; this sequence later underwent thrusting and high-pressure metamorphism in the middle or late Cretaceous.

This juxtaposition of rock bodies displaying jarringly different histories is characteristic of the Franciscan belt. In some areas the contrasting rock bodies

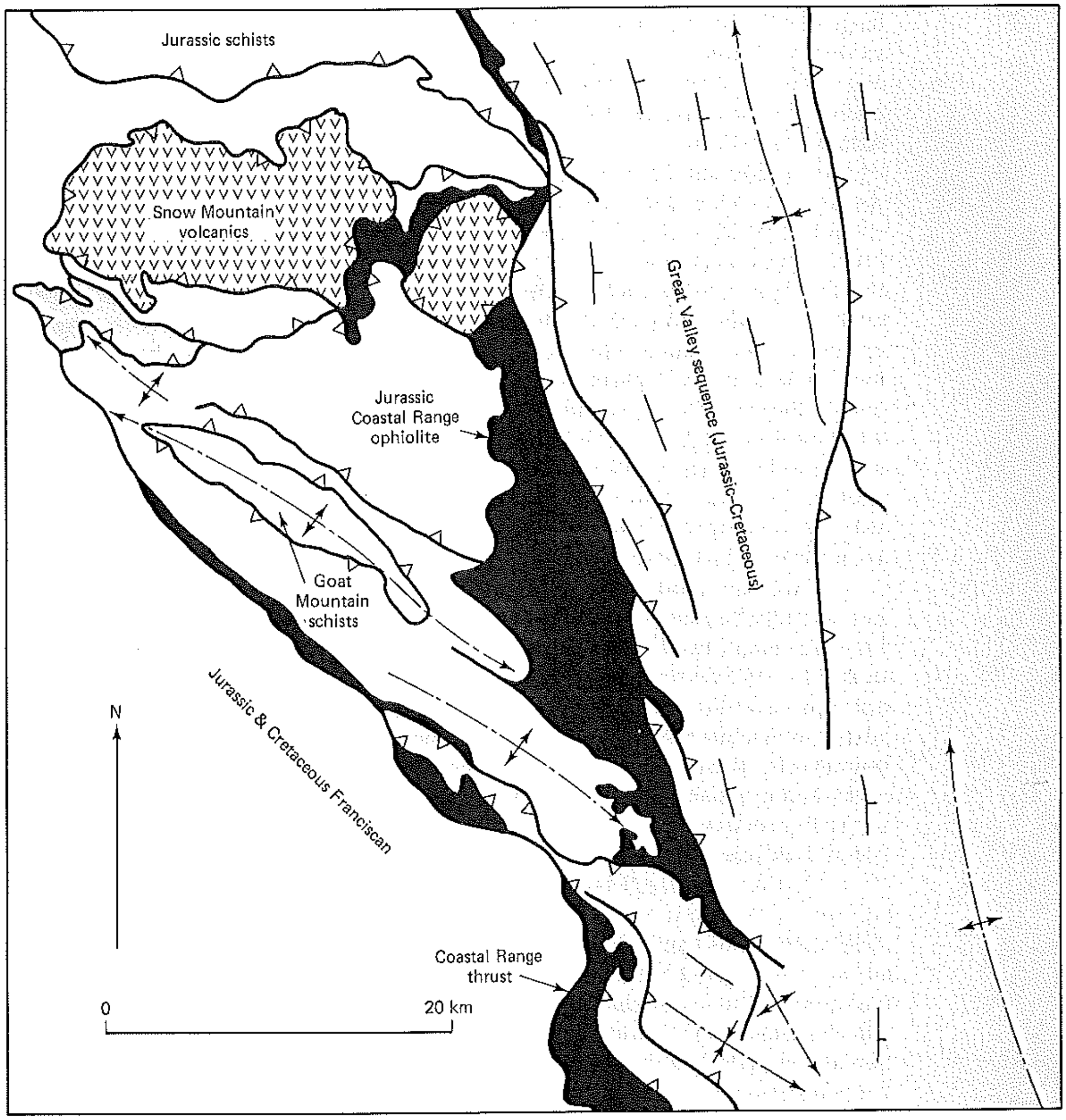

FIGURE 13–14 Map showing typical structure of Coast Range and Great Valley sequence, northern California. (Simplified after Suppe and Foland, 1978. Permission to reproduce granted by the Pacific Section, Society of Economic Paleontologists and Mineralogists.)

are thrust sheets kilometers thick, whereas in other areas they are pieces in the range 1 to 100 m, with the pieces imbedded in a chaotic matrix. For this reason many parts of the Franciscan belt are classified as mélange (Chapter 8). The two best-documented processes by which these fragments of contrasting history are juxtaposed are (1) submarine landslides (olistostromes) and (2) thrust imbrication with multiple episodes of thrusting.

The image we are building up of the Franciscan belt is one that is internally complex and in many ways chaotic, juxtaposing rocks of disparate histories. Nevertheless, it is similarly complex throughout its length and is quite different from adjacent tectonic belts in its assemblage of oceanic rocks, high-pressure metamorphism, and age. The tectonic unity of the Franciscan belt is the sort of unity we might expect of a long-lived accretionary wedge containing an assemblage of diverse oceanic and trench sediments now strongly deformed and subject to varying degrees of metamorphism, depending on their position in the accretionary wedge. Furthermore, we expect many times of metamorphism, deformation, and sedimentation to be represented, rather than a single orogenic event of short duration.

In spite of the general agreement between the structural and tectonic properties of the Franciscan belt and our image of what an accretionary wedge should look like, there are some differences. In particular, the accretionary processes in the Franciscan do not appear to have been monotonous throughout the 150 m.y. between late Jurassic and middle Tertiary; the data on history of tectonism in the Franciscan belt suggest that it contains certain discrete orogenic episodes rather than a fully continuous and uniform accretionary process. If we look at the times of high-pressure metamorphism and sedimentation over a large part of the Franciscan, we see that it is broadly continuous from late Jurassic to late Cretaceous, but on a closer look we see that much of the high-pressure metamorphism is in the range 155 to 135 m.y. (late Jurassic and earliest Cretaceous) and 115 to 95 m.y. (late early Cretaceous to late Cretaceous) with very little after 95 m.y. or between 115 and 135 m.y. Thus the high-pressure Franciscan belt displays some signs of discrete orogenic events.

If we look more carefully at the batholith belt, we find a somewhat similar but not fully in-phase set of magmatic pulses. The late Jurassic period of activity in the magmatic arc (Fig. 13-11) corresponds closely in time with the major period of late Jurassic and earliest Cretaceous metamorphism in the Franciscan belt; this episode is traditionally called the Nevadan orogeny, as discussed above. The early Cretaceous is a time of relative igneous and metamorphic inactivity in the paired belt. The middle to late Cretaceous is again a time of coeval magmatism and metamorphism, the Sevier orogeny. Tectonic activity resumed in the Cenozoic. The correspondence in time between tectonic events in the two halves of the paired belt reinforces the interpretation that they are indeed a genetic pair.

THE GREAT VALLEY SEQUENCE

The Franciscan and batholith basement belts are generally separated by 50 to 100 km along much of their length. The coeval late Mesozoic rocks that fill this space are unmetamorphosed sediments of the *Great Valley sequence* in California and its equivalents to the north and south. These sediments represent a very thick sedimentary basin, exceeding 10 km in some areas, which records an intermediate paleogeography between the magmatic arc on the east and the Franciscan ocean on the west. In terms of analogies with present-day subducting continental margins, the Great Valley sequence apparently fills a forearc basin such as the forearc basin of Guatemala shown in Figure 1-15.

For comparison, a cross section of the Great Valley sequence in northern California is given in Figure 13-15, extending from the Franciscan belt in the Coast Range on the west to the batholith belt in the Sierra Nevada on the east. Note that the Great Valley sequence onlaps the plutonic and metamorphic rocks of the batholith belt so that younger Upper Cretaceous part of the Great Valley sequence overlies Upper Jurassic and Lower Cretaceous plutonic rocks on the western edge of the batholith belt. In a few areas such as Baja California, the eastern edge of the Great Valley sequence actually includes volcaniclastic sediments; furthermore, the Great Valley sequence has a sedimentary petrography that shows a derivation from a volcanic and plutonic arc. Therefore, it is well documented that the Great Valley sequence was in its present position relative to the batholith belt at the time of its deposition.

The relationship of the Great Valley sequence to the coeval Franciscan belt is more complex. As we travel across the Great Valley of California, from the Sierra Nevada to the Coast Ranges (Fig. 13-15), the Great Valley sequence is undeformed and gradually thickens for two-thirds of the way across the valley. Then suddenly at the western edge of the valley (Figs. 13-14 and 13-15), the section turns up on end and, in a single homocline, exposes one of the thickest complete stratigraphic sections you will see anywhere, more than 10 km of Upper Jurassic and Cretaceous clastic sediments, which sits depositionally on a disturbed ophiolite sequence. This ophiolite sequence is called the *Coast Range Ophiolite* (Hopson, Mattinson, and Pessagno, 1981) and is a great sheet of Middle Jurassic (165–155 m.y.) oceanic crust, which extends from Oregon to southern California and probably southern Baja California.

So we see that the Great Valley sequence sits on several different basements. On the east it sits on the batholith belt, whereas on the west it sits on disturbed Middle Jurassic Coast Range Ophiolite; therefore, the base of the Great Valley sequence, which ranges in age from late Jurassic to middle Cretaceous, covers over the contact between continental margin basement and oceanic basement hidden under the western Great Valley. The Coast Range Ophiolite appears to have been attached to North America in the late Jurassic Nevadan orogeny, one of the great orogenic episodes of the Cordillera.

The Coast Range Ophiolite is generally not a fully intact sheet of Jurassic oceanic crust. In most areas it is disturbed with the volcanic and mafic plutonic parts of the lithosphere commonly missing; we generally pass from the basal Great Valley sequence, which contains sediments rich in ophiolitic detritus, directly into mantle rocks, serpentinized peridotites, and dunites. Apparently much of the Coast Range Ophiolite was strongly disturbed before the deposition of the basal Great Valley sequence. In other areas the ophiolitic sequence is fairly complete; for example at Point Sal, along the southern California coast, you can walk along the shore, observing a nearly complete cross section of the oceanic crust and upper mantle in the sea cliffs (Hopson, Mattinson, and Pessagno, 1981).

Immediately below the Coast Range Ophiolite in most areas are the Franciscan rocks, and the contact is generally a thrust fault, the *Coast Range thrust,* which is best exposed along the western edge of the Great Valley in the foothills of the Coast Ranges (Figs. 13-14 and 13-15). The rocks of the Franciscan have the same range in stratigraphic age as the Great Valley sequence, so the Coast Range thrust juxtaposes two coeval sequences, putting the Upper Jurassic and Cretaceous Great Valley sequence and its ophiolitic basement over the Upper Jurassic and Cretaceous Franciscan.

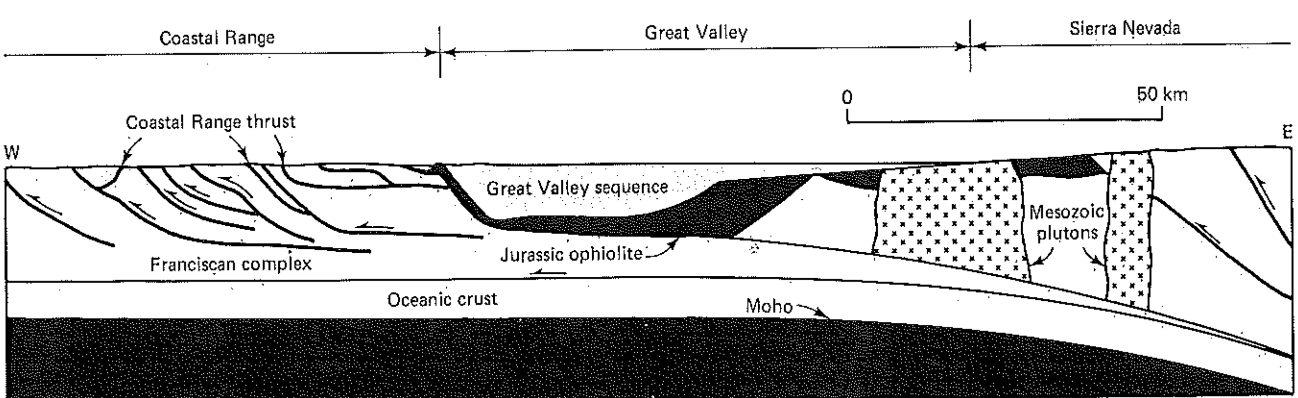

FIGURE 13–15 Cross section of Coastal Range, Great Valley, and Sierra Nevada, northern California. (Based in part on Suppe, 1979.)

The Coast Range thrust has some unusual properties for a thrust fault; in some ways it is quite different from any fault we have encountered heretofore in the Appalachians or Cordilleras. The Franciscan rocks below the thrust have been metamorphosed at various times in the late Jurassic and Cretaceous at very high pressures, equivalent to 20 to 35 km burial, whereas the Great Valley sequence has never been buried deeper than its maximum stratigraphic depth of 10 to 12 km. Therefore, from the point of view of metamorphism, the Coast Range thrust might be mistaken for a normal fault because it seems to cut out section. The cover of the Franciscan terrain at the time of metamorphism is now missing. This strong contrast in history across the Coast Range thrust serves to emphasize how very important this structure is. It has a minimum slip of 100 km, and the actual slip is probably much more because klippen of Great Valley sequence and Coast Range Ophiolite overlie the Franciscan rocks all the way across the Coastal Range for a distance of about 100 km in northern California (Fig. 13-15). Furthermore, the Great Valley sequence and Franciscan rocks are imbricated in a series of Cenozoic thrusts that account for roughly another 100 km of shortening (Figs. 13-14 and 13-15).

In summary, the Great Valley sequence represents an intermediate paleogeography in the late Mesozoic between the batholith belt on the east and the Franciscan belt on the west and apparently represents a forearc basin. Nevertheless, the late Mesozoic boundary between the forearc basin and the Franciscan accretionary wedge is now obscure because of very large displacement on the Coast Range thrust and on thrusts that deform the Coast Range thrust and imbricate the Great Valley sequence and coeval Franciscan metamorphic rocks.

SEVIER FOLD-AND-THRUST BELT

So far we have seen that the late Mesozoic tectonics of the western margin of North America from Oregon to southern Baja California are dominated by a continental-margin subduction system comprising the Franciscan accretionary wedge, Great Valley forearc basin, and batholith-belt magmatic arc. We now move east of the magmatic arc to another coeval zone of deformation, the *Sevier fold-and-thrust belt*, which is part of the great Cordilleran foreland fold-and-thrust belt, extending from Alaska to Central America (Fig. 13-1).

The Sevier fold-and-thrust belt lies at the edge of the North American craton and runs right along the boundary between the Paleozoic miogeosyncline and craton (Figs. 13-9 and 13-16). For this reason the fold-and-thrust belt is not parallel to the paired basement belts to the west. The map pattern of the paired basement belts is controlled by the edge of the continental-margin subduction, whereas the edge of the overthrust belt is controlled by the edge of thick sediments around the orogenic belt.

Directly east of the fold-and-thrust belt is a late Mesozoic foredeep, a sedimentary basin that contains a record of the nearby orogenic activity. For this reason we first look at the foredeep to get an overview of the orogenic history by noting the ages of important clastic wedges. A restored stratigraphic cross section of the western part of the foredeep is shown in Figure 13-17. Three big tongues of clastic sediments are apparent: (1) late Jurassic–earliest Cretaceous, (2) late early Cretaceous to early late Cretaceous, and (3) middle late Cretaceous. These times of orogenic sedimentation show that mountain building is broadly coeval between the paired basement belt and the overthrust belt; indeed the time of lessening activity in the middle early Cretaceous seems to be present in both.

You have probably noticed that we have generally dated orogenic activity in our discussions of the Appalachian and Cordillera tectonics using the ages of

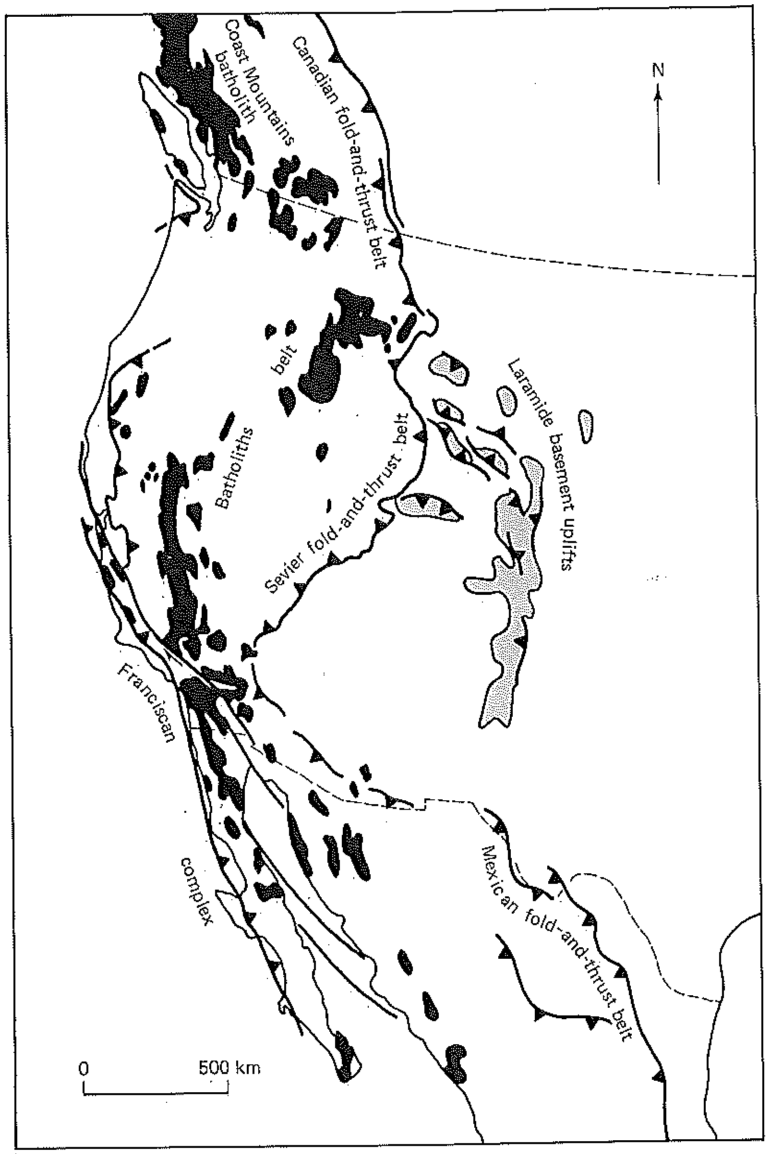

FIGURE 13–16 Map of major late Mesozoic and early Cenozoic structural belts, including the Franciscan and related accretionary complexes, the batholith belt (in black), the fold-and-thrust belt, and the Laramide basement uplifts.

clastic wedges, metamorphism, and igneous activity. Well-constrained ages of actual structures such as faults and folds are much more difficult to obtain and therefore often play a subsidiary role in discussions of tectonic history. The Sevier overthrust belt is an exception to this generality in that some of the important structures are reasonably well dated based on angular unconformities and distinctive detritus eroded from the thrust sheets.

An example of one of these well-dated structures is the great Absaroka thrust of Wyoming (Fig. 13-18). The older branch contains an overturned fault-propagation fold of Paleozoic through Lower Cretaceous strata in the upper plate, which overrides middle Santonian (Upper Cretaceous) conglomerates. These conglomerates were apparently eroded from the advancing thrust sheet because it contains detritus similar to formations in the thrust sheet, with rock types that do not resist abrasion very well. The conglomerate is very coarse, with some boulders exceeding 2 m. As you pass up through the conglomeratic sequence, clasts of older and older formations appear, indicating that uplift and denudation was going on during deposition.

A second branch of the Absaroka thrust was activated to the east of the early Absaroka thrust after the middle Santonian conglomerate was deposited and

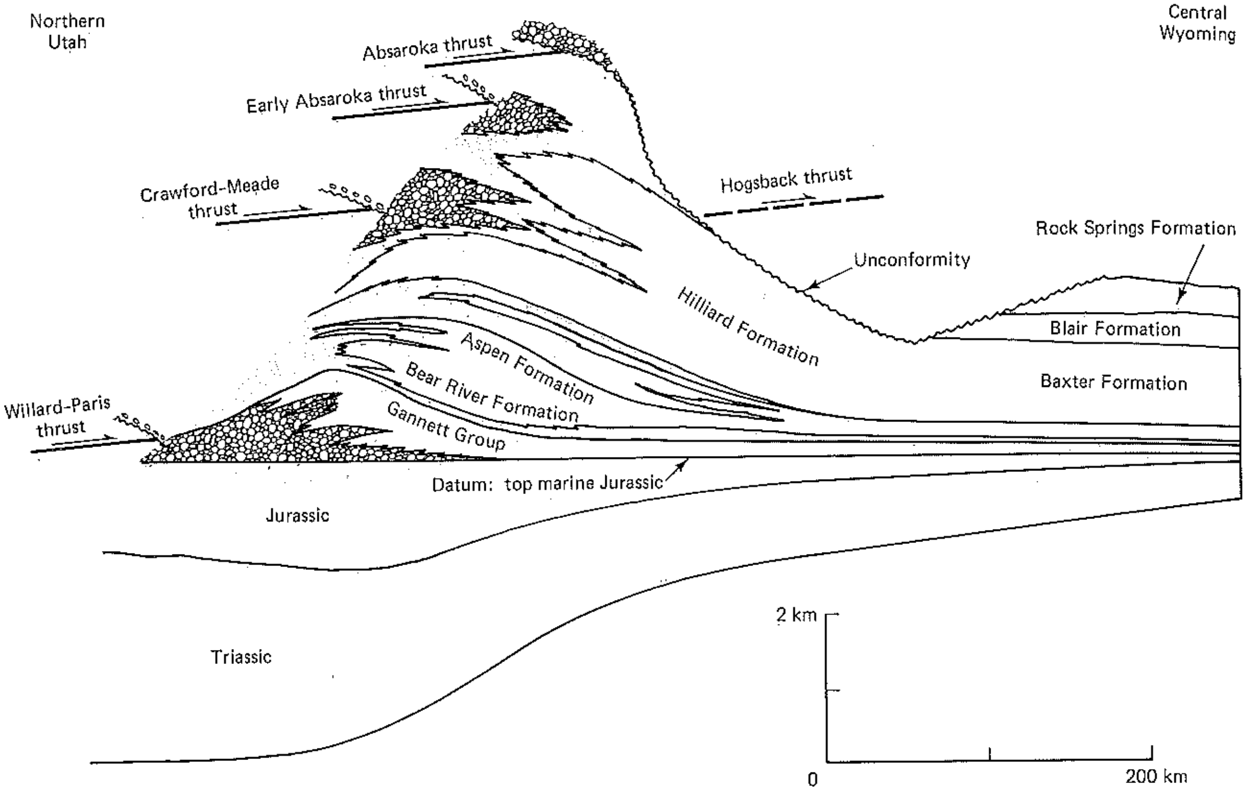

FIGURE 13–17 Restored stratigraphic diagram of mesozoic foredeep deposits in western Wyoming and northern Utah, showing relationship between clastic wedges and thrusting. (Simplified after Royse, Warner, and Reese, 1975.)

overridden. The entire region is eroded and overlain by the Lower Tertiary Evanston Formation, which places a lower limit on the youngest deformation. Similar field and subsurface relations allow fairly precise dating of many of

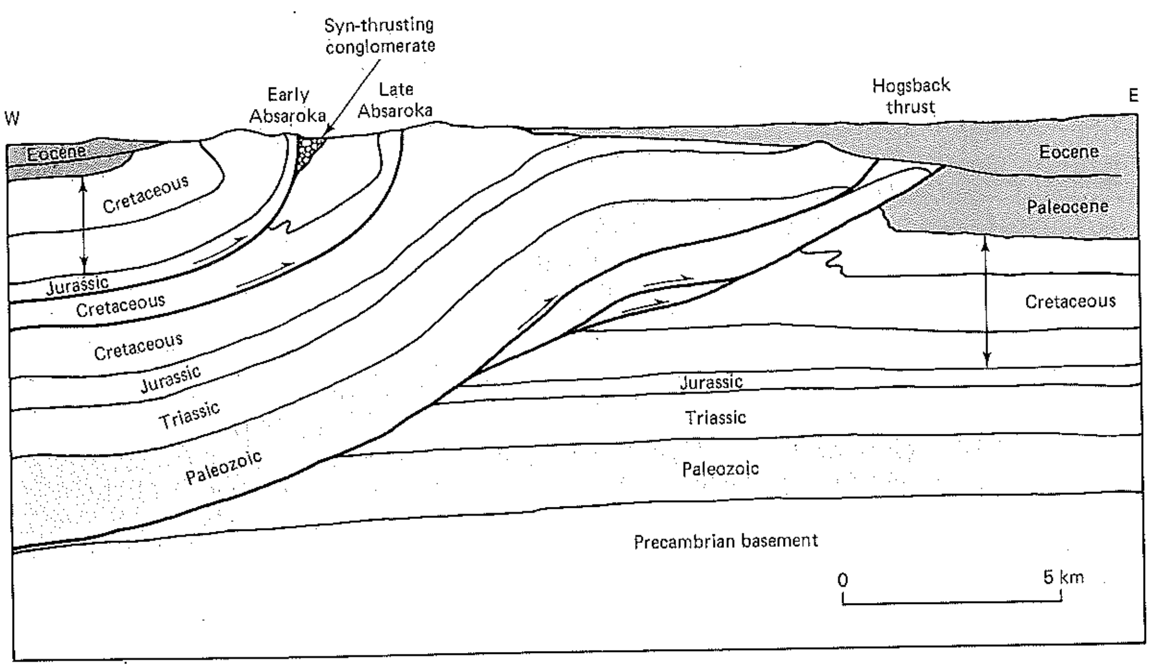

FIGURE 13–18 Cross section of early and late Absaroka thrusts and Hogsback thrust, Sevier fold-and-thrust belt, Wyoming. (After Royse, Warner, and Reese, 1975).

the major structures of the Sevier fold-and-thrust belt and show that deformation generally coincides with the times of the clastic wedges of the foredeep (Fig. 13-17).

SOUTHERN CANADIAN FOLD-AND-THRUST BELT

This brief introduction to the regional structural geology and tectonics of the Cordillera has focused almost exclusively on the segment in the western United States because it is regionally the best known and geographically the most-commonly visited. However, the best-known and best-displayed segment of the Cordilleran fold-and-thrust belt is not in the United States, but in the southern Canadian Rocky Mountains in western Alberta and eastern British Columbia. Here the structure is displayed in a grand alpine fashion on the mountain faces of Banff and Jasper National Parks (Fig. 9-6). For example, Figure 13-19 shows some complex imbricate fold-and-thrust structure in the Boule Range along the Athabaska River east of Jasper.

The structure of the foreland of the Canadian Rocky Mountains is naturally divided into a series of belts running parallel to the mountain front, each characterized by a distinctive structural style, distinctive stratigraphy, and distinctive topography. These belts may be seen in the cross section of the Canadian Rockies in Figure 13-20 and are the Alberta basin (Fig. 1-8), Foothills, Front Ranges, Main Ranges, Purcell anticlinorium, and Kootenay arc. These belts and the cross section take us about halfway across the 700-km-wide Cordilleran mountain belt in southern Canada (Fig. 13-16).

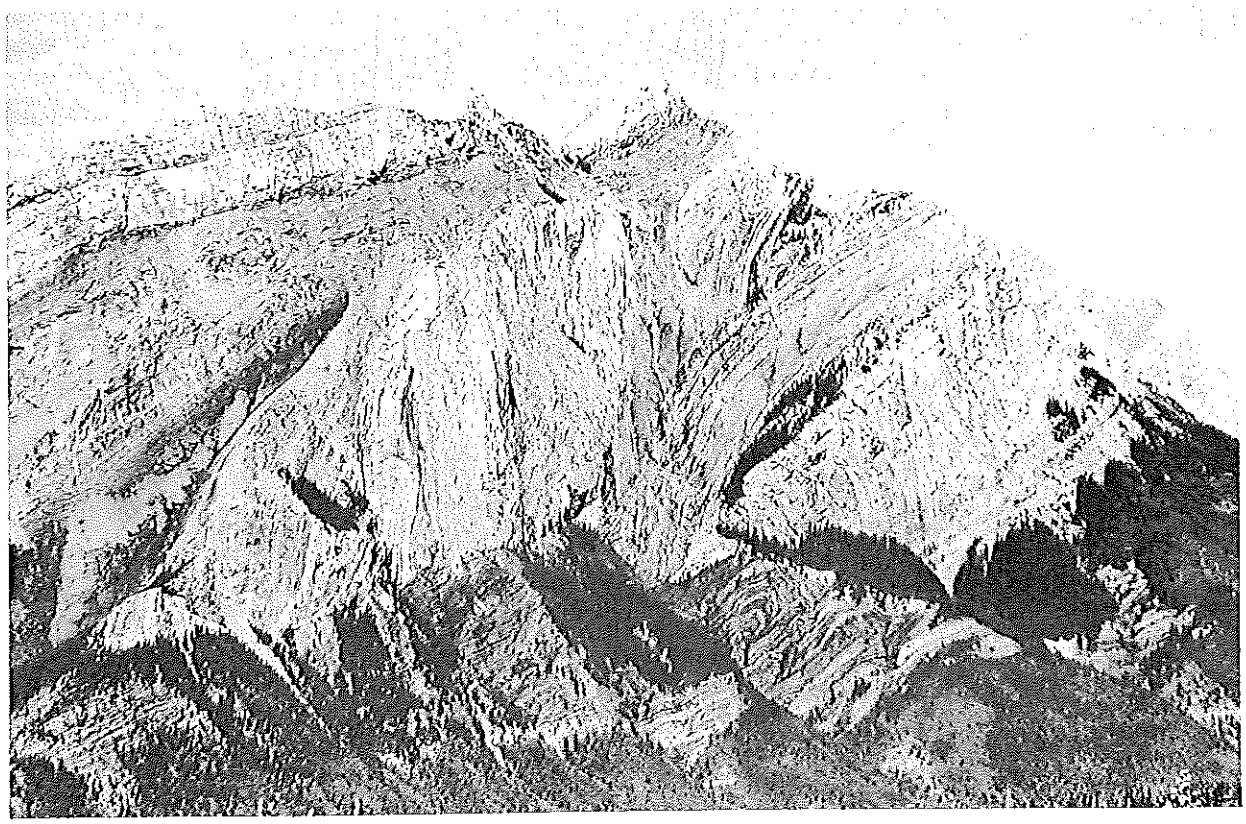

FIGURE 13–19 Frontal thrusts of the Front Ranges, Athabaska River, Jasper National Park, Canada. The foreground is underlain by Mesozoic clastic sediments of the Foothills belt, whereas the mountain exposes Upper Paleozoic carbonates of the Front Ranges.

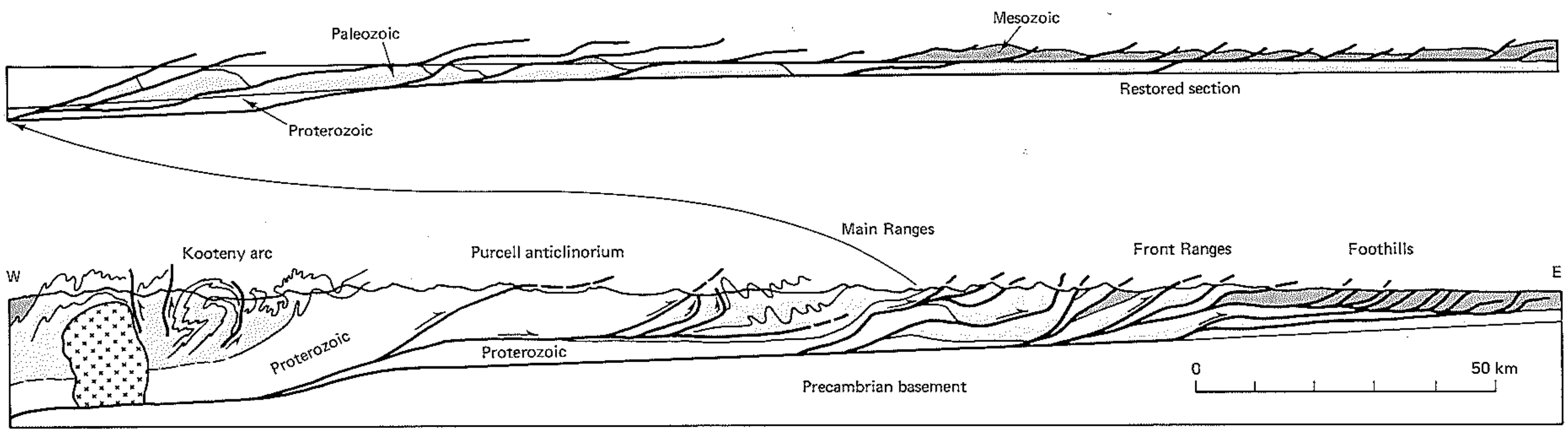

FIGURE 13–20 Cross section and restored section of the eastern Canadian Rockies. (Simplified after Price and Fermor, undated.)

The easternmost belt is the *Alberta basin,* which is the undeformed foredeep in front of the fold-and-thrust belt (Fig. 1-8). This basin and its deformed equivalents to the west contain as much as 4 km of Upper Jurassic through Lower Tertiary clastic sediments derived from the deforming Cordillera to the west. The times of orogenic activity recorded in the southern Canadian foredeep are not entirely the same as those recorded on the eastern margin of the Sevier orogenic belt in the United States (Fig. 13-17). Both are active in the late Jurassic and Cretaceous, but the fold-and-thrust belt in the western United States ceases just before the end of the Cretaceous, whereas the overthrust belt in southern Canada and Montana continued to deform through the Paleocene and into the Eocene. These differences in history will become important in the next section, when we discuss the latest Cretaceous and early Tertiary Laramide orogeny.

The *Foothills belt* of the Rocky Mountains is characterized by low topographic relief and very strong imbricate thrusting and folding of the same foredeep section that is undeformed in the Alberta basin to the east. The Foothills belt is in essence the deformed western margin of the Alberta basin. The structure is well known because of extensive exploration for petroleum, with the initial discoveries in the 1920s. One of the important contributions of this exploration to structural geology is the discovery that the strong deformation exposed at the surface, displaying steep dips, abundant fault repetitions of the stratigraphy, and tight folding, does not extend to great depth. At depths of 3 to 5 km, we pass into undeformed Paleozoic cratonic sediments, which overlie Precambrian crystalline basement (Fig. 13-21). Most of the thrust faults flatten to décollement horizons in the Mesozoic or, in some places, the upper Paleozoic section. The horizontal shortening above the basal décollement is very large; the 50-km-wide Foothills belt in Figure 13-20 was originally about 100 km wide based on retrodeformation of the cross section.

The western limit of the Foothills belt is marked by the abrupt topographic front that is the beginning of the *Front Ranges*. The Rocky Mountain front, such as that east of Jasper (Fig. 13-19), generally marks the first thrust sheet with Paleozoic rocks that rises to the surface. The thrust that serves as the frontal thrust changes along strike in *en échelon* arrangement, but each places cliff-forming Cambrian and younger carbonates over the easily erodable Cretaceous sandstones and shales. These frontal thrusts include some of the best-known thrust faults in North American geology: the Lewis thrust in Glacier-Waterton National Park (Fig. 8-26), the McConnell thrust in Banff National Park (Fig. 8-4), and the Boule thrust at Jasper National Park (Fig. 13-19).

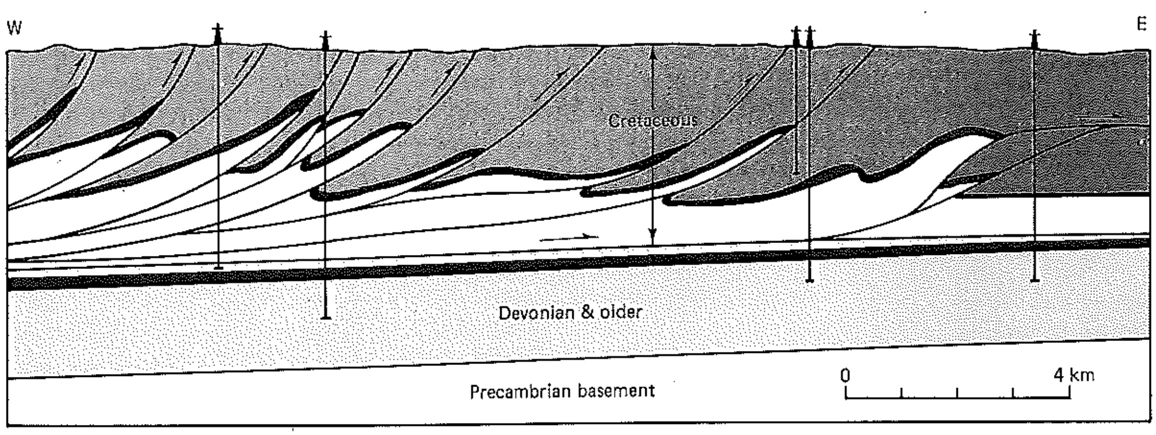

FIGURE 13–21 Detailed cross section of part of Foothills belt of the Canadian fold-and-thrust belt, showing complex imbrication of the Mesozoic strata but underformed Paleozoic at depth. (Simplified after Ollerenshaw, 1973.)

The Paleozoic and lower Mesozoic carbonate section that forms the imbricated thrust structure of the Front Ranges (Fig. 13-20) displays geosynclinal thicknesses of 4 to 10 km, thickening to the west with stratigraphy that is very similar to the Cordilleran miogeosyncline in the western United States (for example, Fig. 13-4). In contrast, the coeval section that is present in the subsurface of the Foothills Belt and Alberta basin is cratonic in character and only 1 to 1.5 km thick. Thus the stratigraphic change from Paleozoic craton to geosynclinal continental margin is now structurally compressed and displaced and represented by the Foothills–Front Range structural boundary (Fig. 13-20).

The major thrust sheets in the Front Ranges are much more widely spaced than those in the Foothills Belt—about 5 km, as opposed to 1 to 2 km (Figs. 13-21 and 8-4). This difference apparently reflects the much-larger steps in décollement that are characteristic of the geosynclinal section of the Front Ranges; the thrusts generally step from low in the Cambrian to the Upper Paleozoic or Lower Mesozoic in a single step. Each of the mountain ranges or ridges running parallel to strike in the Front Ranges is generally a single thrust sheet, with the ridge held up by Paleozoic cliff-forming carbonates. As we travel across the Front Ranges—for example, along the Trans-Canadian Highway (Fig. 8-4)—we see the same Paleozoic stratigraphic section repeated in each mountain range with the strata always dipping west. This is the first and most-obvious indication of thrust imbrication. Only after much mapping is it found that major thrust faults generally

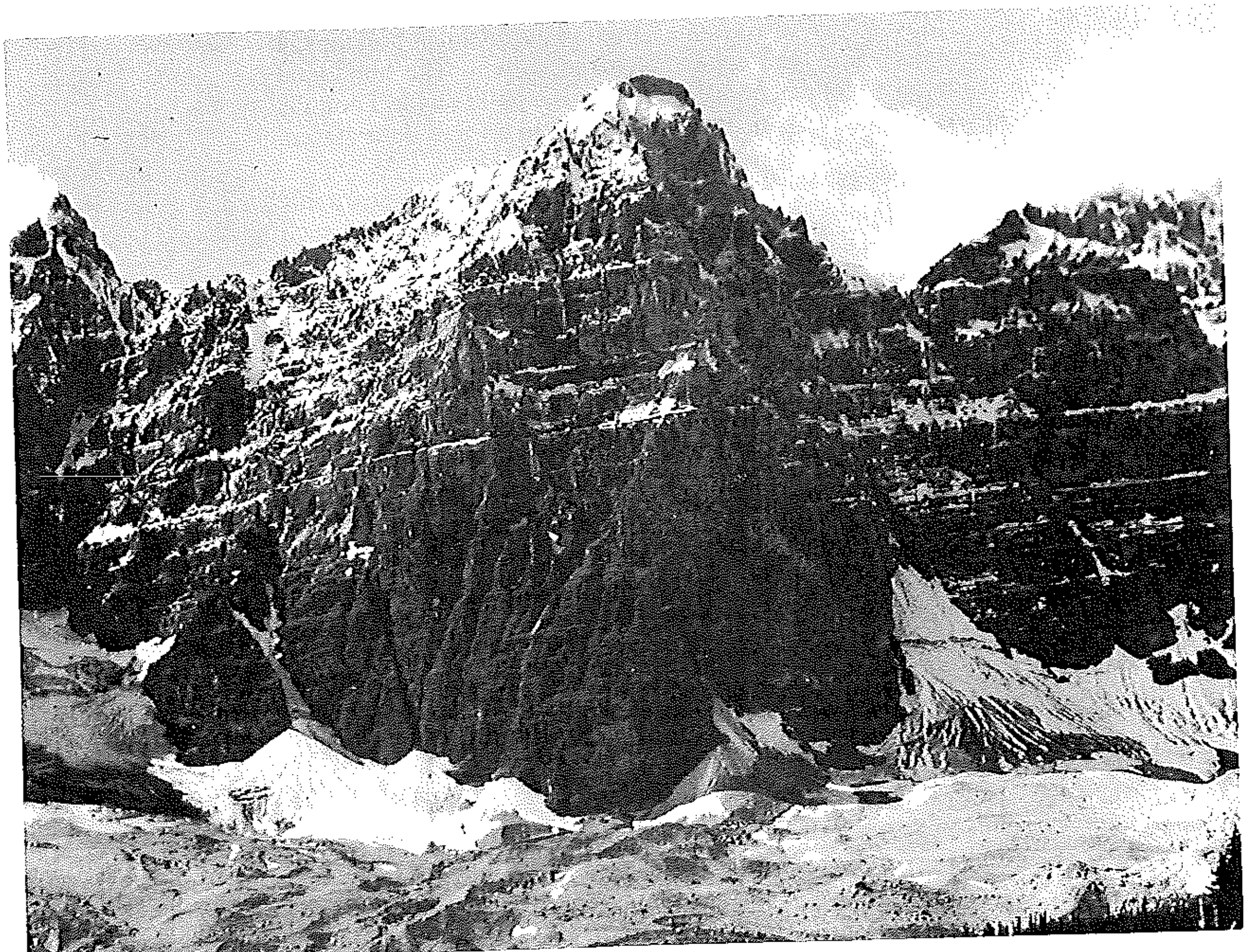

FIGURE 13–22 Little-deformed Cambrian carbonates of the eastern Main Ranges, near Lake Louise, Banff National Park. Compare with Figure 13-23.

run along the parallel valleys, hidden by the thick forest cover. The décollement structure of the Front Ranges is substantiated by outcrops of the thrusts here and there, seismic reflection profiling, deep drilling, and retrodeformable cross sections. The shortening of the 50-km-wide Front Ranges in Figure 13-20 is about 75 km.

As we travel through the western Front Ranges and into the Main Ranges, we encounter two important changes in the Paleozoic and late Precambrian continental-margin stratigraphy that bring about major changes in the mountain structure. First, the basal décollement steps down in the section several kilometers into a thick section of late Precambrian clastic sediments, which is correlative with the late Precambrian sediments of the western United States already discussed. The main structural effect of the appearance of Upper Precambrian strata below the Cambrian is much thicker thrust sheets (Fig. 13-20).

A second, more-structurally profound, stratigraphic change is caused by a sedimentary facies change in the Cambrian. In the east the Cambrian is characterized by thick massive carbonate sequences, typical of the early Paleozoic carbonate bank. This stratigraphy has deformed stiffly, forming great thrust sheets, but displaying little deformation on the scale of a mountain face (Fig. 13-22). As we move west into the western Main Ranges, this Cambrian carbonate bank abruptly terminates and drops into a deeper-water shale basin containing carbonate submarine landslide breccias derived from the bank. This Cambrian shale sequence is much less resistant to buckling than the stiffer bank carbonates; therefore, they display very tight chevron folds that dominate the mountain faces (Fig. 13-23, see also Fig. 13-20).

FIGURE 13–23 Chevron folds of Cambrian shales of the western Main Ranges. Compare with Figure 13-22. (Geological Survey of Canada, Ottawa.)

At this point we truncate our traverse, although we are still less than halfway across the Canadian Cordillera. The rocks have reached only low greenschist facies of regional metamorphism and the shales are deformed to slates. If we were to continue westward (Fig. 13-20), we would pass into high-grade metamorphic rocks, still apparently sediments of the North American continental margin; but eventually we would pass through ophiolitic belts, recording destroyed oceanic basins, and into orogenic terrains that are far-traveled with respect to North America. These *displaced terrains* play an important role in Cordilleran tectonics, which is discussed below, but it is first appropriate to describe the latest Cretaceous and early Tertiary foreland deformation in the United States east of the Sevier fold-and-thrust belt.

LARAMIDE OROGENY (Latest Cretaceous and Early Tertiary)

Latest Cretaceous is a time of tectonic reorganization, particularly in the segment of the Cordillera within the United States. At this time the Franciscan and Batholith belts cease their active metamorphism and magmatism; furthermore, the Sevier fold-and-thrust belt becomes inactive. Orogenic deformation moved east of the Sevier overthrust belt, producing fault-bounded uplifts of Precambrian basement in the formerly stable craton. The latest Cretaceous through Eocene *Laramide orogeny* is named for one of these basement uplifts, the Laramie Range of eastern Wyoming (Fig. 13-24). In contrast, the magmatic arc and fold-and-thrust belt continued to be quite active during Laramide time to the north in the Canadian Rockies and Montana and to the south in Mexico.

The locus of latest Cretaceous through Eocene foreland deformation in the Cordillera has a remarkable map pattern (Fig. 13-16). The two active segments of the fold-and-thrust belt deform completely different stratigraphic sections. The northern segment, in the eastern Canadian Cordillera, deforms the westward-thickening Paleozoic continental margin of the Proto-Pacific ocean in great east-vergent thrust sheets (Fig. 13-20). In contrast, the Mexican fold-and-thrust belt of the Sierra Madre Oriental deforms the eastward-thickening Mesozoic continental margin of the Gulf of Mexico in great east-vergent thrust sheets. In essence the Mexican fold-and-thrust belt lies at the eastern edge of the continent and thrusts oceanward, whereas the Canadian fold-and-thrust belt lies at the western edge of the continent and thrusts cratonward. Between eastern Mexico and western Canada, the locus of Laramide foreland deformation cuts across the North American craton, giving rise to the distinctive fault-bounded uplifts of Precambrian basement.

The Laramide basement uplifts are generally 20 to 30 km across and as a group form an arcuate pattern changing from an east-west strike in the northwest to a north-south strike in the southeast. The margins of most ranges are bounded by thrust faults or sharp fault-related flexures of the sedimentary cover. The flexure on the east side of the Front Range near Denver, Colorado, is shown in Figure 13-25(a); faults of minor displacement exist within the structure, as documented by drilling, but the deformation of the sedimentary cover is accomplished largely by folding at higher levels.

The major underlying structural mechanism of the basement uplifts appears to be thrust faulting. For example, low-angle thrusting on the north side of the Uinta Mountains uplift, documented by seismic data and drilling, has displaced the Precambrian basement approximately 10 km over the adjacent sedimentary basin (Fig. 13-25(b)). Most of the structural relief of the Uinta uplift is accomplished by slip along the bounding thrust.

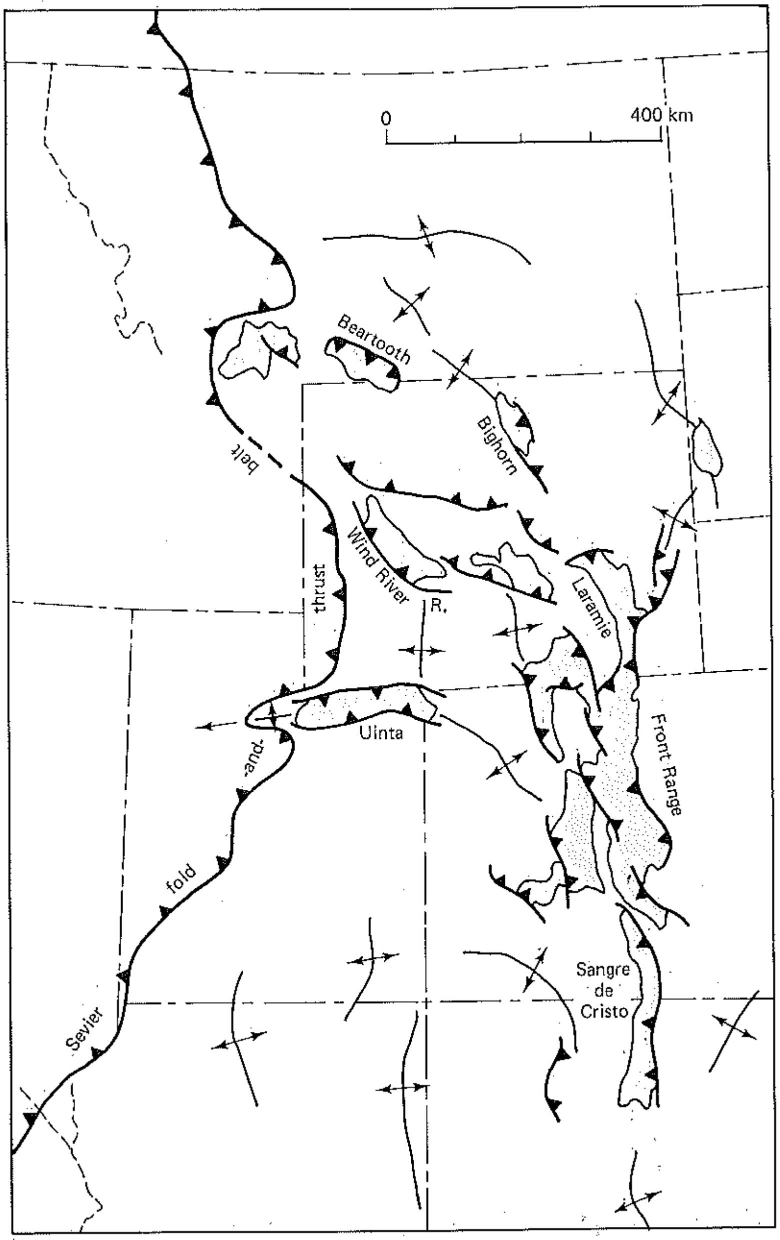

FIGURE 13–24 Map of main structural features of Laramide Rocky Mountains. (Simplified after Hamilton, 1978. Permission to reproduce granted by the Pacific Section, Society of Economic Paleontologists and Mineralogists.)

Similar bounding thrust faults have been documented for most of the major Laramide uplifts, the best known of which is the Wind River uplift in western Wyoming (Fig. 13-26). The Wind River uplift is asymmetric with a gentle homocline of the sedimentary cover on the northeast side and a major thrust fault on the southwest side, the *Wind River thrust,* which places the Precambrian basement over sediments of the flanking Green River basin. The Wind River thrust has been traced to a depth of about 20 km by seismic-reflection profiling and appears to flatten in the lower crust or along the Moho, based on considerations of retrodeformability (Fig. 13-26). The structural mechanism of uplift is slip on the Wind River thrust.

DISPLACED TERRAINS

Let us summarize the broadest features of Cordilleran tectonics as far as we have gone—up to the early Tertiary—in terms of just two contrasting regimes. (1) For

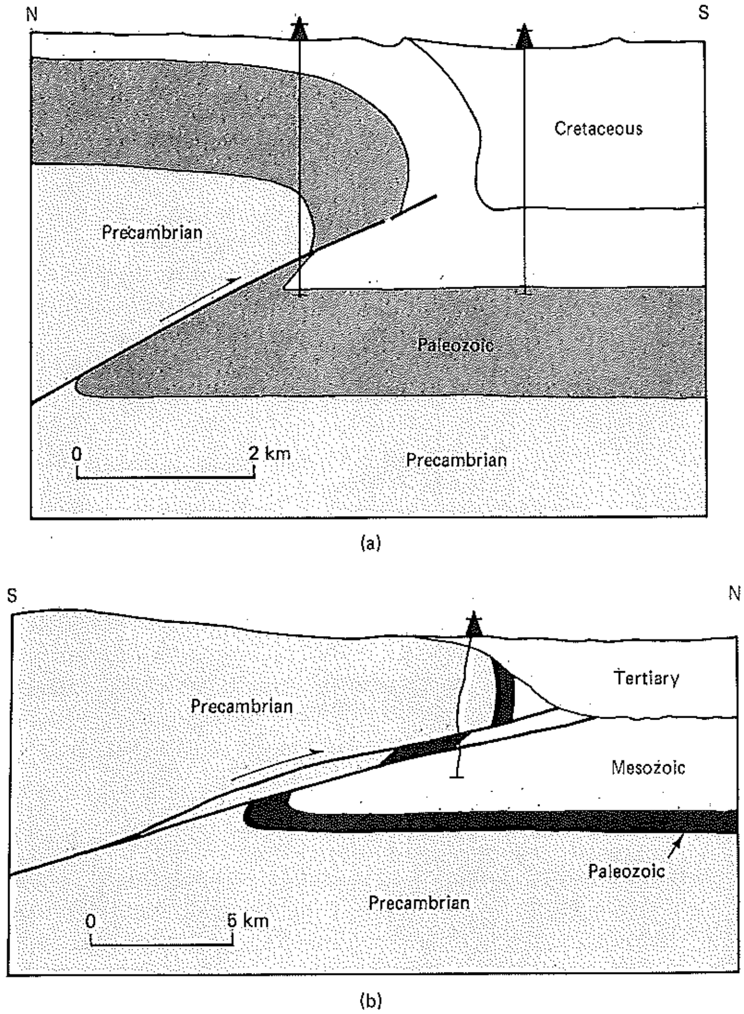

FIGURE 13–25 Structures of Laramide Rocky Mountains. (a) Willow Creek thrust, Colorado. (Modified after Berg, 1962.) (b) Uinta Mountains. (Simplified from Gries, 1983.)

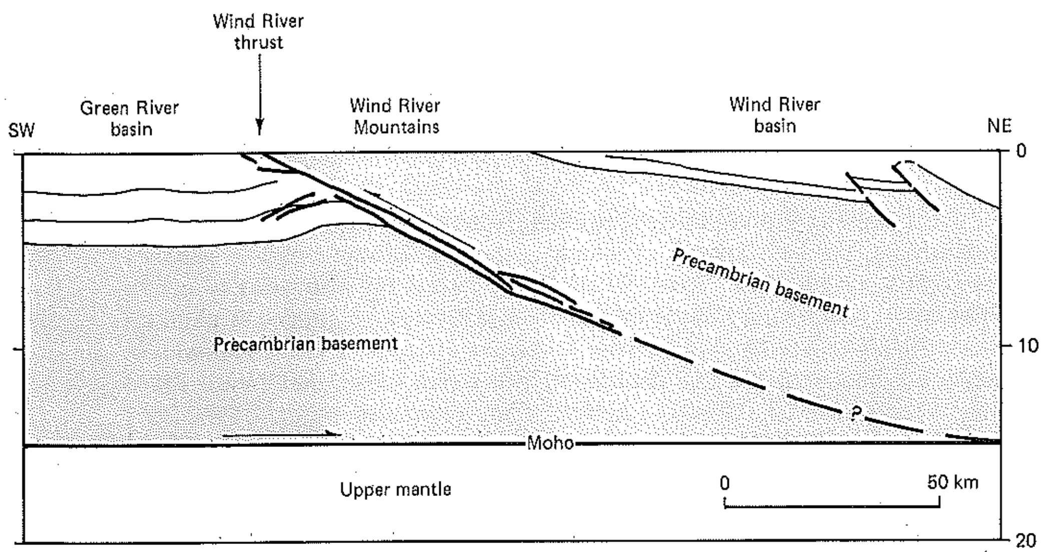

FIGURE 13–26 Cross section of Wind River thrust, based on seismic line. (*Note:* Vertical dimension is time.) (Modified after Brewer and others, 1979.)

the half-billion years beginning with late Proterozoic rifting (approximately 850 m.y. ago), western North America was the site of a stable continental shelf, slope, and rise bordering the Proto-Pacific ocean. (2) This stable tectonic regime began to change with the late Devonian and early Carboniferous Antler orogeny. By late Triassic or early Jurassic, a regime of continental-margin subduction of the Indonesian type was well established, consuming the vast late Proterozoic and Paleozoic Proto-Pacific ocean and eventually much Mesozoic and Cenozoic ocean as well.

It follows that all Paleozoic rocks in the Cordillera found west of the original Paleozoic continental margin must be far-traveled and are by some process accreted to North America. Furthermore, much Mesozoic and Cenozoic rock, such as the Franciscan belt and Coast Range Ophiolite, appears to be accreted. In all, over half of the Cordilleran mountain belt is known or suspected to be far-traveled and accreted to the edge of North America during the regime of plate convergence. The western half of the Cordillera is a vast collage of fault-bounded accreted terrains together with superjacent post-accretion sediments, such as the Great Valley sequence, and intrusive and extrusive rocks of the continental-margin magmatic arc.

More than fifty discrete *displaced terrains* have been recognized in the western Cordillera; a few of the larger ones are shown in Figure 13-27. Each displaced terrain is characterized by a distinctive internal stratigraphy, structural style, and geologic history that is foreign to North America and to the other terrains, with boundaries that cannot be interpreted as facies changes or unconformities but must be fundamentally faults of great displacement. Geologic understanding of many of these displaced terrains is limited, but enough are well known that we can provide important examples and make certain generalizations that provide us with a picture of this fundamental aspect of Cordilleran tectonics.

Displaced terrains naturally divide themselves into two fundamental types: (1) *accretionary terrains*, such as the Franciscan belt, produced by the piecewise imbrication of many small, fault-bounded fragments and (2) *stratigraphic terrains* that contain a coherent stratigraphic sequence and substantial geologic history that predates accretion of the terrain to North America as major fragments. Stratigraphic terrains appear to include island arcs, aseismic ridges, oceanic plateaus, and small fragments of probable continental crust. Two of the major stratigraphic terrains are discussed in the following paragraphs: Stikinia and Wrangellia (Fig. 13-27).

As we travel westward across the Canadian fold-and-thrust belt, we encounter Paleozoic rocks that record paleogeographies progressively farther offshore of the North American craton (Fig. 13-20). Eventually, we encounter displaced terrains exotic to North America. One of the most-obvious of these is the long, linear *Cache Creek terrain* of early Carboniferous through late Triassic age, approximately 1400 km long and up to 75 km wide (Fig. 13-27). The Cache Creek belt is a structurally complex accretionary terrain composed of radiolarian chert, argillite, submarine basaltic volcanic rocks, gabbros, and ultramafic rocks. This assemblage of oceanic rock types indicates that the Cache Creek terrain represents the remains of a late Paleozoic and early Mesozoic Proto-Pacific ocean.

Just west of the Cache Creek accretionary terrain is a major late Paleozoic and Mesozoic stratigraphic terrain, the *Stikine terrain* of northwestern and north central British Columbia (Fig. 13-27). The Stikine terrain in Carboniferous time is composed of stratigraphy indicative of an island arc: marine and subaerial basalts, andesites, and rhyolites, with associated volcaniclastic sediments and shallow-

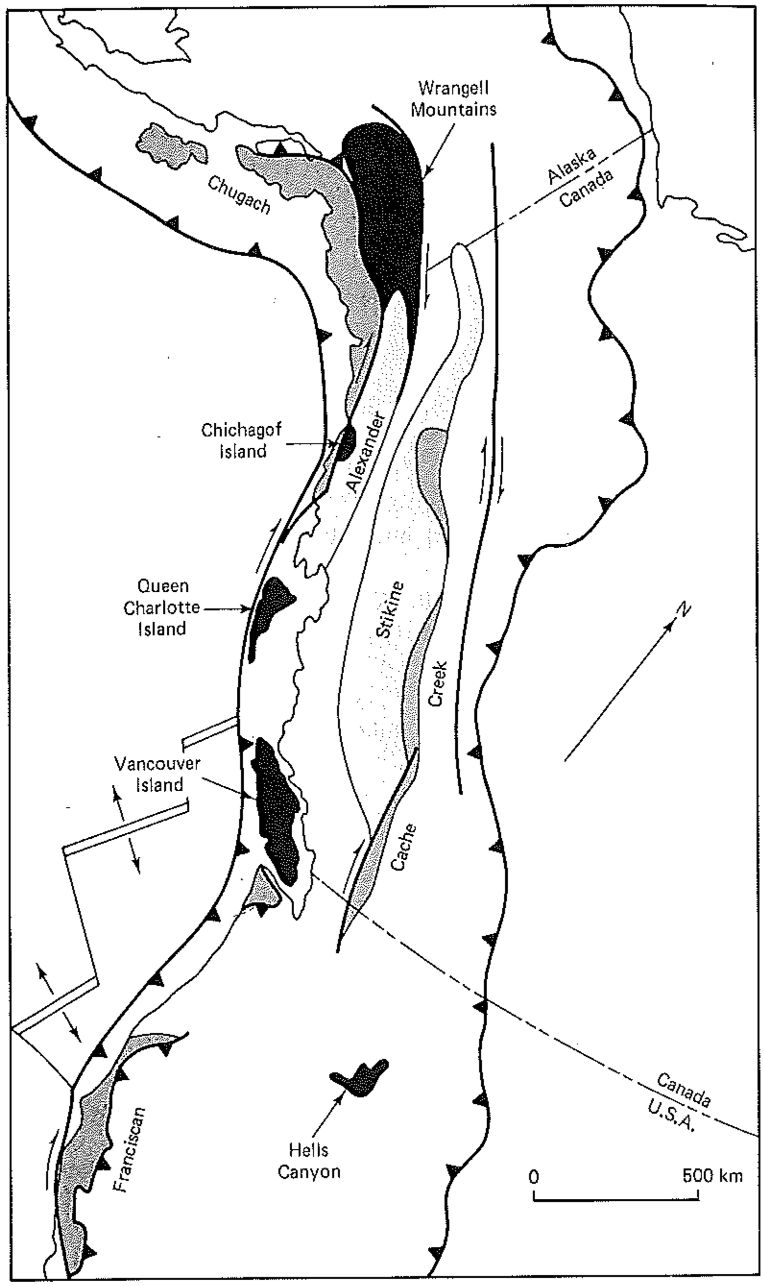

FIGURE 13–27 Map showing some of the major displaced terranes in the Cordilleran mountain belt. (Compiled from maps of Coney, Jones, and Monger, 1980, and Jones, Silberling, and Hillhouse, 1977.)

water limestones. Paleomagnetic data from overlying Lower Mesozoic volcanogenic sediments indicates a paleolatitude about 15° south of the present position of the Stikine terrain. Paleolatitude is determined from the inclination of the magnetization vector relative to the originally horizontal bedding surface (Chapter 2, Problem 2-8). The northward component of post–early Jurassic motion of the Stikine terrain relative to North America is therefore about 1300 km; the amount of longitudinal displacement is unknown. The stratigraphy of the Cache Creek and Stikine terrains seems to require that they be exotic to North America and the paleomagnetic measurements provide confirmation of substantial motion.

West of the Stikine terrain and west of the plutonic and high-grade metamorphic rocks of the Coast Mountains batholith (Figs. 13-16 and 13-27) are a set of widely dispersed fragments of the *Wrangellia terrain,* noted for a very distinctive sequence of Middle and Upper Triassic rocks, as well as underlying Upper Paleozoic rocks. Fragments of Wrangellia extend nearly 3000 km from the Wrangell Mountains of southern Alaska to the Hells Canyon of eastern Oregon (Figs. 13-27 and 13-28). The characteristic stratigraphic feature of Wrangellia is a thick 3 to 6 km sequence of Middle and Upper Triassic marine tholeiitic flood basalts, called the Karmutsen Formation on Vancouver and Queen Charlotte Islands and the Nikolai Greenstone in Alaska, overlain disconformably by Upper Triassic shallow-water limestones (Fig. 13-28). It is remarkable that such similar stratigraphy exists over a distance of 3000 km.

Wrangellia does not represent an early Mesozoic ophiolitic sequence, because where the bases of the Triassic basalts are exposed, they are underlain by Upper Paleozoic limestones, argillites, and volcanic rocks of island-arc affinities. Apparently the flood basalts poured out over a preexisting island-arc terrain and are perhaps analogous to the submerged basaltic platforms or hot-spot traces that exist today in the western Pacific and Indian Oceans.

Paleomagnetic data from the Triassic flood basalts indicate that Wrangellia is far-traveled. Data from both the Wrangell Mountains and Vancouver Island give consistently low paleolatitudes, 15° north or south of the Triassic equator, depending on the polarity epoch.

There are a number of other significant displaced terrains for which large differences in stratigraphy and paleogeography are documented. Paleomagnetic data and faunal evidence indicate significant northward motion relative to North America, similar to that shown by Stikinia and Wrangellia. It is less clear how much eastward motion accompanied the displacement of these terrains; for example, did they sail in from the central or western Proto-Pacific ocean? Some

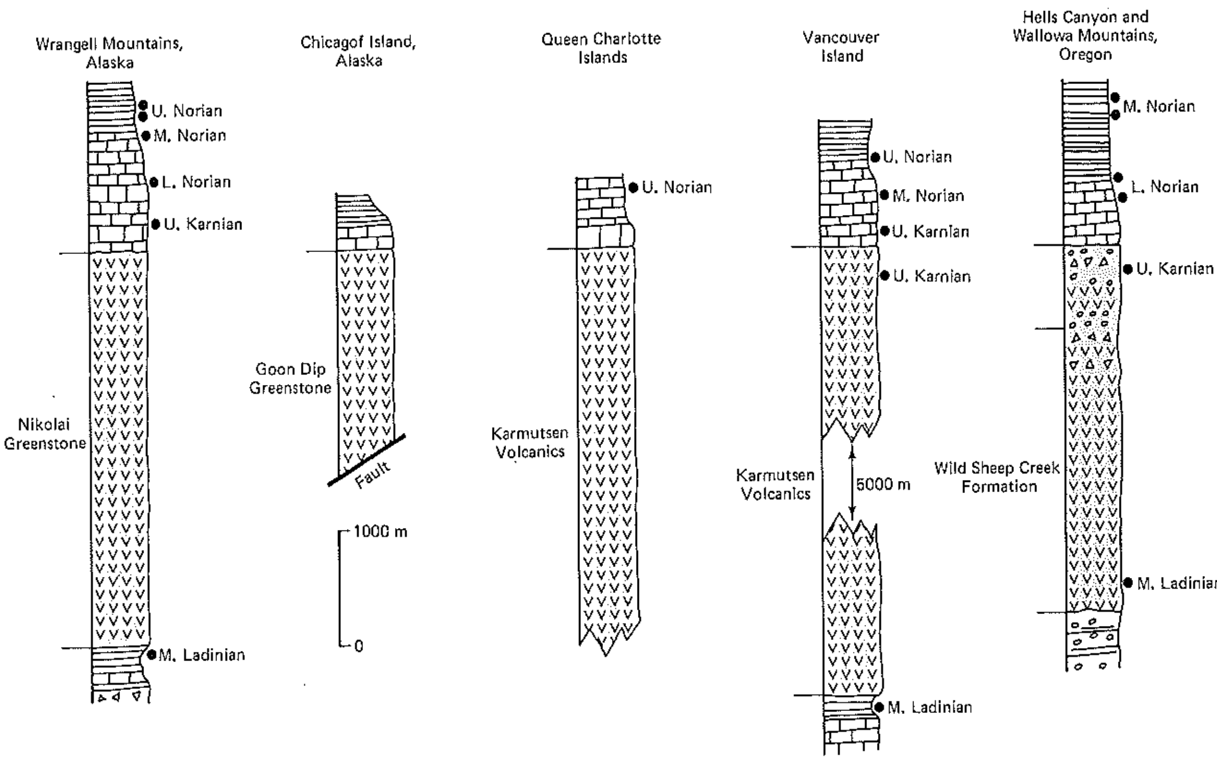

FIGURE 13–28 Comparison of stratigraphies of various parts of Wrangellia between Idaho and Alaska. See Figure 13-27 for locations. (After Jones, Silberling, and Hillhouse. Reproduced by permission of the National Research Council of Canada from the Canadian Journal of Earth Sciences, v. 14, 1977.)

eastward motion is required by the closing of the Cache Creek ocean basin. Furthermore, not all northward motion took place prior to accretion of these displaced terrains to North America. A great system of right-lateral strike-slip faults exists along the Cordilleran mountain belt from Baja California to Alaska (Figs. 13-2 and 13-29). Even today the Wrangell Mountains are moving northwestward along the Denali fault.

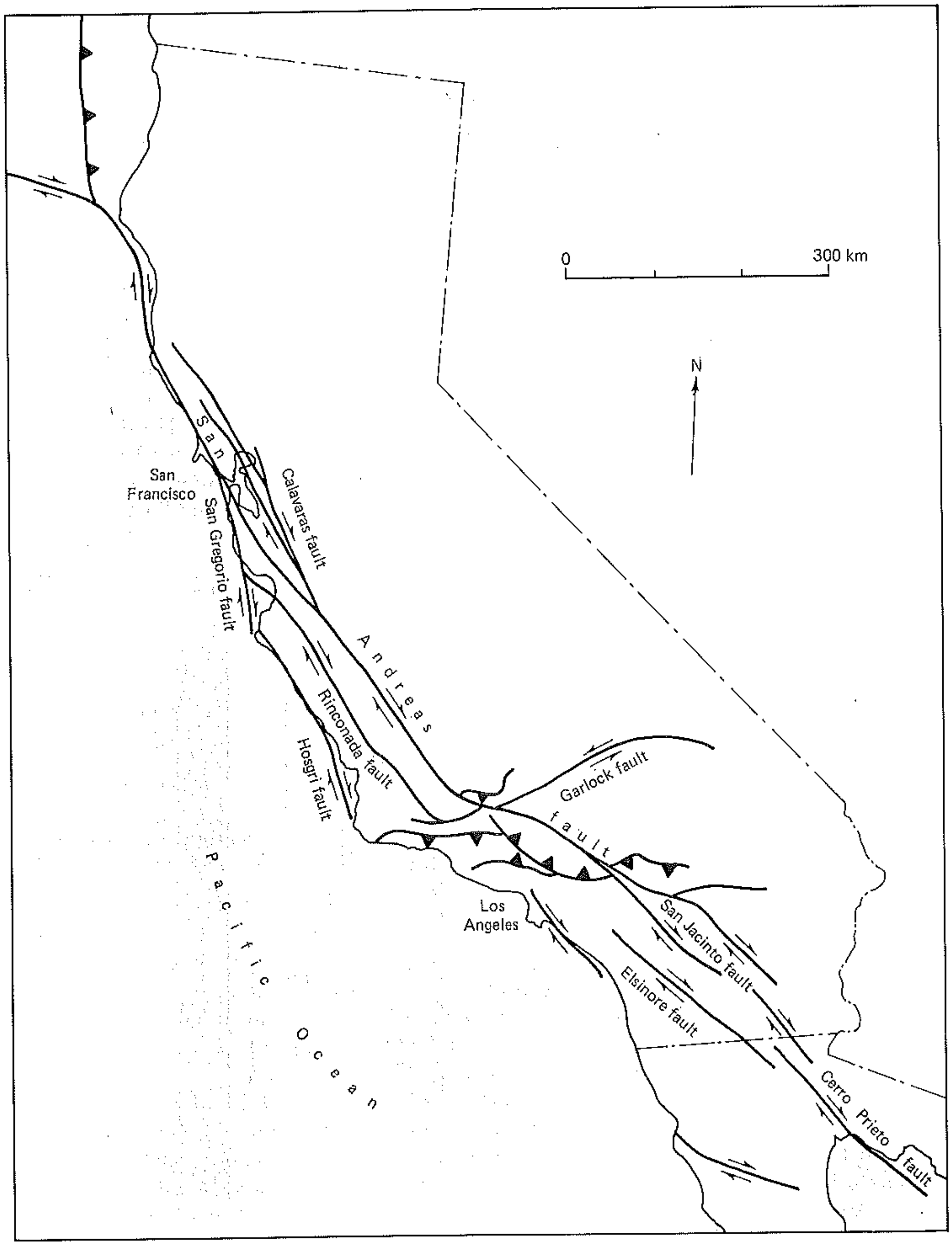

FIGURE 13–29 Major strike-slip faults of the San Andreas system, California.

STRIKE-SLIP ALONG THE PACIFIC MARGIN: The San Andreas Fault

The large northward displacements of Wrangellia and other terrains that are suggested by paleomagnetic evidence are among the more-recent of a whole series of discoveries of Mesozoic to Recent northward displacements along the western margin of North America. The first important indication of northward displacement was the great 1906 San Francisco earthquake (Chapter 8), which showed that an active right-lateral strike-slip fault zone, the *San Andreas fault system,* runs the length of California (Figure 13-29).

At first it seemed to many people that the San Andreas fault might have only a few kilometers of slip, but as regional geologic mapping of California became more and more complete, large regional mismatches in geology across the fault were discovered (Fig. 13-30). In 1953 M. L. Hill and T. L. Dibblee showed regional mapping evidence that suggested displacements of hundreds of kilometers with the displacements becoming larger with age. The late Mesozoic Franciscan-Batholith pair of basement belts were apparently offset by about 500 km (Fig. 13-13).

The idea of such large slip was not immediately accepted by many people and became very controversial. There were two principal reasons for the controversy. First, large horizontal displacements, including both large-scale thrusting and continental drift, were very unpopular at the time among the more articulate North American geologists. Secondly, there existed some apparently conflicting evidence. Some areas displayed apparent matches in strata across the fault, whereas other areas displayed strong mismatches. Geologists were naturally impressed with the evidence most familiar to their personal experience, and the San Andreas fault is so long (1200 km) that most geologists were familiar with at most small segments. T. L. Dibblee played a very special role in the discovery of large slip on the San Andreas fault because he was intimately familiar with much of its length. He was a tall, thin, quiet man, a member of an old California pioneer

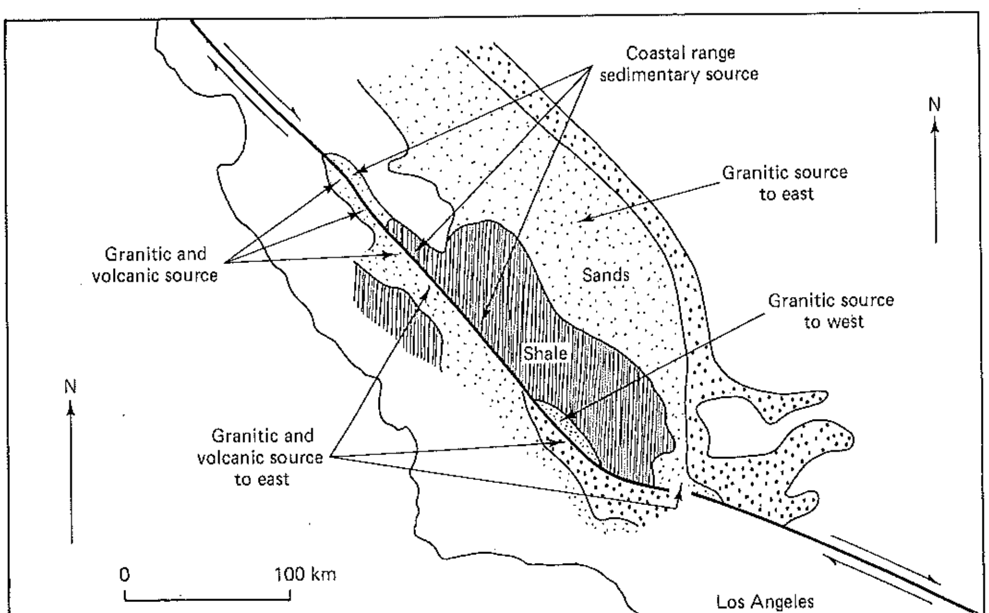

FIGURE 13–30 Discordant Upper Miocene sedimentary facies across the San Andreas fault in central California. (After Huffman, 1972).

family, who loved to spend his time in the field mapping. In the course of his career he mapped several hundred geologic quadrangles in California, many of them along or near the San Andreas fault. Dibblee's experience provided the detailed, but regional, perspective that forced people to consider seriously the uncomfortable possibility of very large slip.

By now parts of the displacement history have been very well documented and the idea of large slip is no longer controversial. For example, Figure 13-30 is a map of central California showing the mismatch in Upper Miocene sedimentary facies and potential sediment-source terrains across the San Andreas fault. There

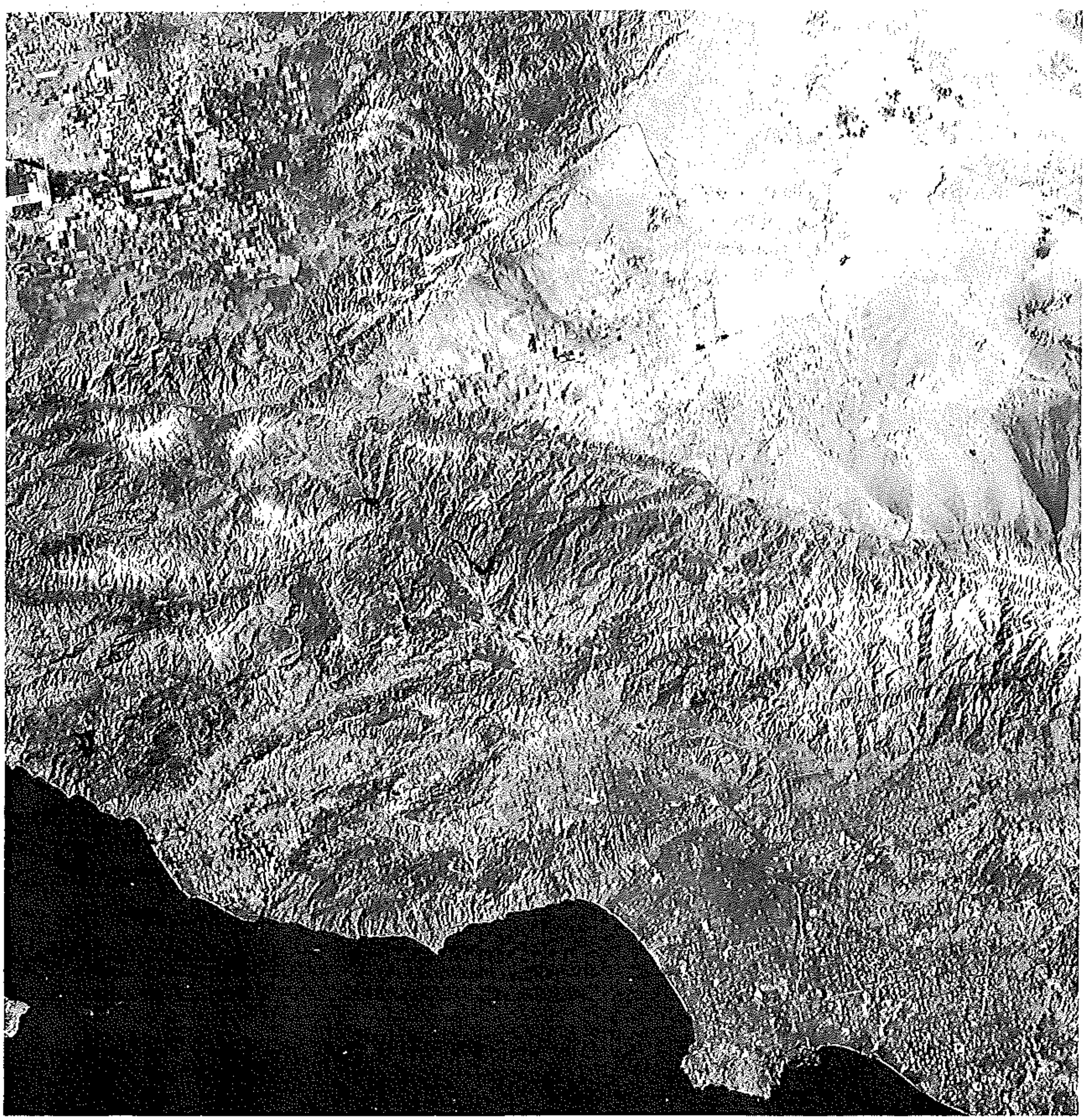

FIGURE 13–31 Satellite image of part of Transverse Ranges. The large, light region in the upper right is the Mojave block, bounded on the south (bottom) by the San Andreas fault and on the northwest by the Garlock fault. Los Angeles is in the lower right.

are obvious mismatches along about half of the fault length. Along the other half there are sediments that might appear superficially similar, but a close examination of their sedimentary petrography and source directions show them also to be mismatched. By reversing the right slip by about 235 km, the Upper Miocene (10 m.y.) geology appears to match quite well.

Similar evidence has been used to match geology across the fault at other times. Earlier Miocene (22 m.y.), Oligocene, and Eocene strata all show a 300-km offset; therefore, in central California the San Andreas fault appears to have been inactive in early to middle Tertiary times and began slipping in late Miocene. Some evidence, such as the offset Mesozoic paired basement belts, hints at a period of strike slip during Laramide time. There is also evidence that suggests or documents important Neogene strike slip in the range 25 to 100 km on a half-dozen other faults, subparallel to the San Andreas (Crowell, 1979).

The principal exception to the pattern of the late Neogene right slip along the San Andreas fault system is the left-lateral *Garlock fault,* which strikes about 80° to the regional trend of the San Andreas (Fig. 13-29). The Garlock fault joins the San Andreas fault in the most-complex part of the San Andreas system, the Transverse Ranges of southern California (Figs. 8-23 and 13-31). A variety of aspects of the Mesozoic through Precambrian geology appear to match across the Garlock fault if it has about 65 km of left-lateral slip. The most striking is a distinctive Mesozoic diabase dike swarm (Fig. 13-32).

This apparent slip of 65 km on the Garlock fault is large relative to the total length of the fault, 260 km. Similarly, the 250 to 500 km of slip on the San Andreas fault is large relative to its length of about 1200 km. Where does the slip go?

The slip on Garlock fault at its eastern end appears to go into the system of normal faults that produce the grabens between the Sierra Nevada and Death

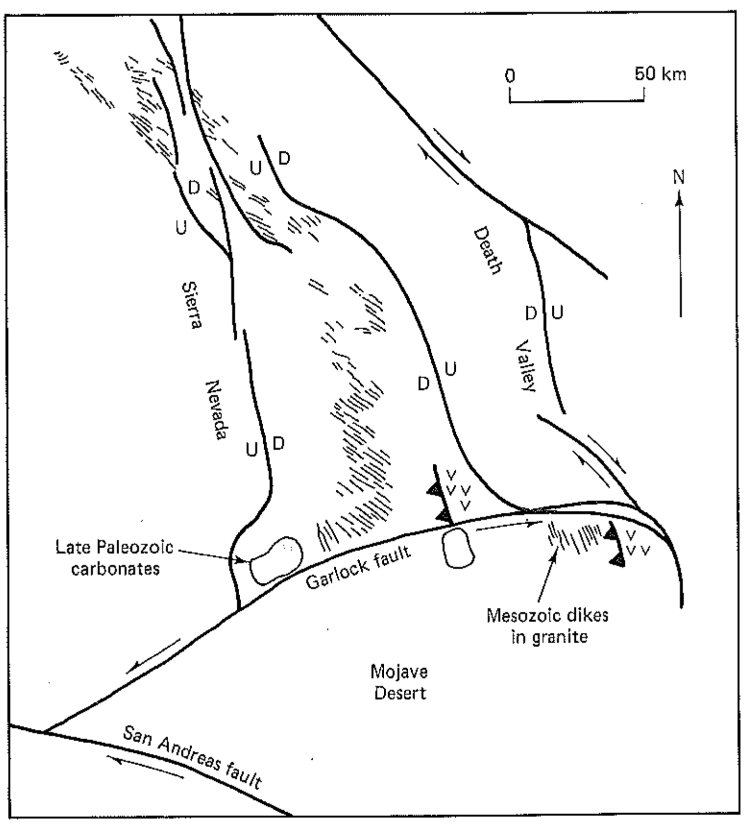

FIGURE 13–32 Offset structures along the Garlock fault, southern California. (Compiled from Smith, 1962, and Davis and Burchfiel, 1973.)

FIGURE 13–33 Oblique aerial view to the west of normal fault scarps of the eastern face of the Sierra Nevada, Owens Valley graben, California. Mount Whitney, the highest peak in the continental United States (4418 m.), is in the center skyline. Ancient shorelines of Owens Lake are in foreground. (U.S. Geological Survey.)

Valley and strike perpendicular to the Garlock fault. One of these active grabens is Owen Valley, just east of the Sierra Nevada, shown in Figure 13-33. The slip of the Garlock fault at its western end appears to deform the San Andreas system in a complex way, contributing to the great bend in the San Andreas in southern California (Fig. 13-29). The Garlock fault appears to be a transform or tear fault at the southern end of a zone of extension east of the Sierra Nevada.

Most of the 240-km slip on the San Andreas fault in the Pliocene and Pleistocene (post 4.5 m.y.) passes at its southern end into the Gulf of California rift, which shows an equivalent extension (Figs. 13-2 and 13-29). Other slip on the San Andreas system may pass into zones of extension that exist in the continental margin of southern California and northern Baja California.

The San Andreas system also displays subsidiary zones of extension and compression associated with irregularities of the fault system. Some of these subsidiary structures were described in Chapter 8 (Fig. 8-23). The most-important irregularity is the great bend in southern California (Figs. 8-23(d) and 13-29), which places this segment in compression, producing strong topographic and structural relief, thrust faulting, and folding. This irregularity is the Transverse Ranges, which juxtapose 3-km-high peaks immediately against the sea-level city of Los Angeles (Figure 13-31). The deformation within the Transverse Ranges appears to divide itself into faults dominated by thrust motion and faults dominated by strike-slip (Figure 13-29). The eastern and central Transverse Ranges are thrust-and-reverse-fault bounded sheets and blocks of Precambrian and Mesozoic basement rocks of the Batholith belt. One of these thrusts slipped in the 1971 San Fernando (Los Angeles) earthquake (Fig. 8-2). Farther west, toward the continental margin, compressive structure of the Transverse Ranges is dominated by folding and décollement thrusting of the Great Valley sequence, Franciscan belt, and their Cenozoic sedimentary cover (for example, Fig. 13-34).

PLATE MOTIONS AND CENOZOIC TECTONICS

The present-day configuration of plate boundaries along the western margin of North America, already discussed at the beginning of the chapter (Fig. 13-1), is marked by transform (strike-slip) boundaries between Pacific and North American plates in the two areas where these plates are in contact: (1) western Canada and (2) western California and northern Mexico. In contrast, the subducting boundaries of North America with their associated magmatic arcs exist where oceanic plates intervene between the Pacific and North American plates: (1) central and southern Mexico and Central America, where the Cocos plate intervenes, and (2) Oregon and Washington, where the Gorda plate intervenes. Subduction tectonics also occur along the North American–Pacific boundary in the Gulf of Alaska and the Aleutians because of the east-west orientation. Therefore, many of the first-order present-day tectonic features of the Cordilleran mountain system, such as the San Andreas and Queen Charlotte strike-slip fault

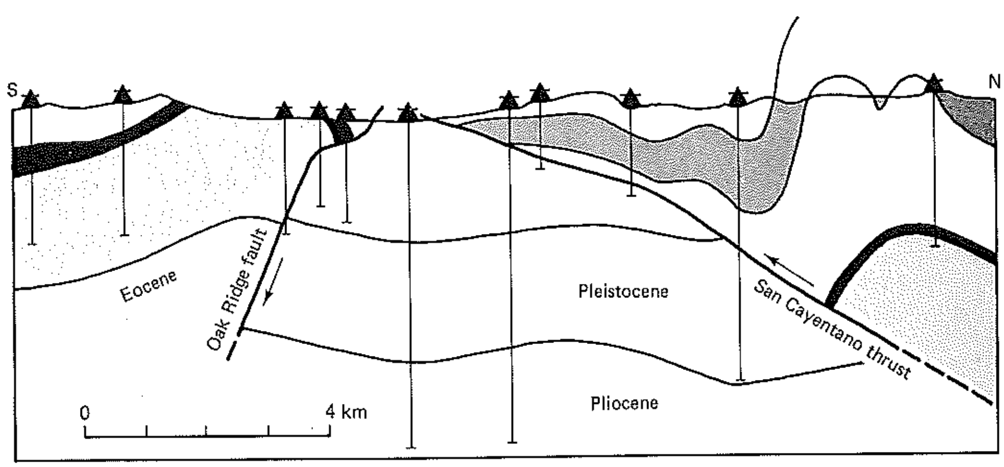

FIGURE 13–34 Cross section of compressive Pleistocene structure within the western Transverse Ranges, California. Oak Ridge fault is a folded old normal fault, now operating as a reverse fault. (After Yeats, J. Geophys. Research, v. 88, p. 569–583, 1983, copyrighted by the American Geophysical Union.)

systems and the Aleutian, Cascade, Mexico–Central American magmatic arcs, have a well-defined setting within the present-day pattern of plate interactions.

In this section we trace the patterns of plate interactions back in time, using primarily the constraints of sea-floor spreading magnetic anomalies, and see how the predicted plate interactions compare with the geologic record in the Cordillera. The most-important observation in making predictions of plate interactions is the fact that the magnetic anomalies of the Pacific plate become progressively older towards the west (Fig. 13-35). Therefore, we can reconstruct the shape of the ridge and transform boundaries of the Pacific plate for various times in the past. We see that the ridge-transform system was formerly much more extensive. A single oceanic plate, the Farallon plate, intervened between the Pacific and North American plates prior to about 30 m.y. ago. The Gorda and Cocos plates represent residual fragments of the single, much more extensive Farallon plate. The strike-slip boundary between the Pacific and North American plates did not exist in California 30 m.y. ago; the Farallon plate intervened.

These observations hold important implications for Cenozoic tectonics in the Cordillera. If sea-floor spreading was symmetric, as it usually is, every bit of the Pacific plate that was created along the Pacific-Farallon ridge system had an associated matching crust in the Farallon plate. This great expanse of Farallon crust is now missing and must have been subducted along a more extensive subduction zone along the western Cordillera; the North America–Gorda and North America–Cocos subduction zones are remnants of this more-extensive system. The area of equivalent crust on the Pacific plate (Fig. 13-35) shows us the minimum amount of Farallon crust that has been subducted.

The relative motions of the Pacific and Farallon (Gorda and Cocos) plates can be completely determined from the marine magnetic anomalies of the eastern Pacific. However, a complete reconstruction of Pacific–Farallon–North American plate interactions requires that we also know the Pacific–North American motions or Farallon–North American motions. This is difficult because of intervening subduction zones during much of the Cenozoic. The Pacific–North American motions for the last 4.5 m.y., during the time that the Gulf of California has been opening, are given by magnetic anomalies at the mouth of the Gulf of California. Prior to 4.5 m.y. (early Pliocene), the motions are only indirectly constrained;

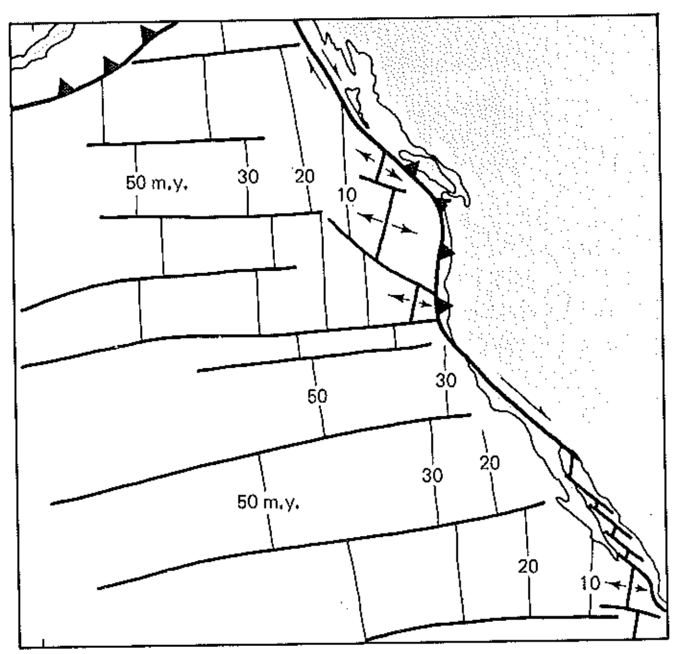

FIGURE 13–35 Map of magnetic anomalies offshore of western North America. Note that the anomalies generally become older toward the west, implying a large amount of subduction along the Cordilleran margin of North America. (Simplified after Atwater, 1970.)

nevertheless, the general pattern of plate interactions is clear and illustrated in Figure 13-36. The key point to remember is that the Pacific plate has been moving north relative to North America through much of the Cenozoic, so that strike-slip tectonics dominated the Cordilleran margin wherever and whenever the Pacific plate was in contact with North America.

According to the magnetic anomalies along the California coast (Fig. 13-35), the Pacific plate began to come in contact with North America no later than about 30 m.y. (Oligocene). This point of contact between the Pacific and North American plates divided the intervening Farallon plate into two isolated parts which are equivalent to the present-day Gorda and Cocos plates. The zone of strike slip has grown progressively (Fig. 13-36) until it is now about 2600 km long. The total Oligocene and younger strike slip predicted from plate motions is about 1000 km, whereas only a fraction of this has taken place along the San Andreas fault because the plate boundary has been jumping into the North American continent by steps. The latest and best-known step is the jumping into the Gulf of California about 4.5 m.y. After this jump, all the 5.5 to 6 cm/year slip between the Pacific and North American plates had been taken up in the Gulf of California and along the San Andreas fault system. Prior to 4.5 m.y., the slip seems to have been taken up on various faults farther to the west in the continent, such as the San Gregorio–Hosgri fault in central California (Fig. 13-29), and possibly initially

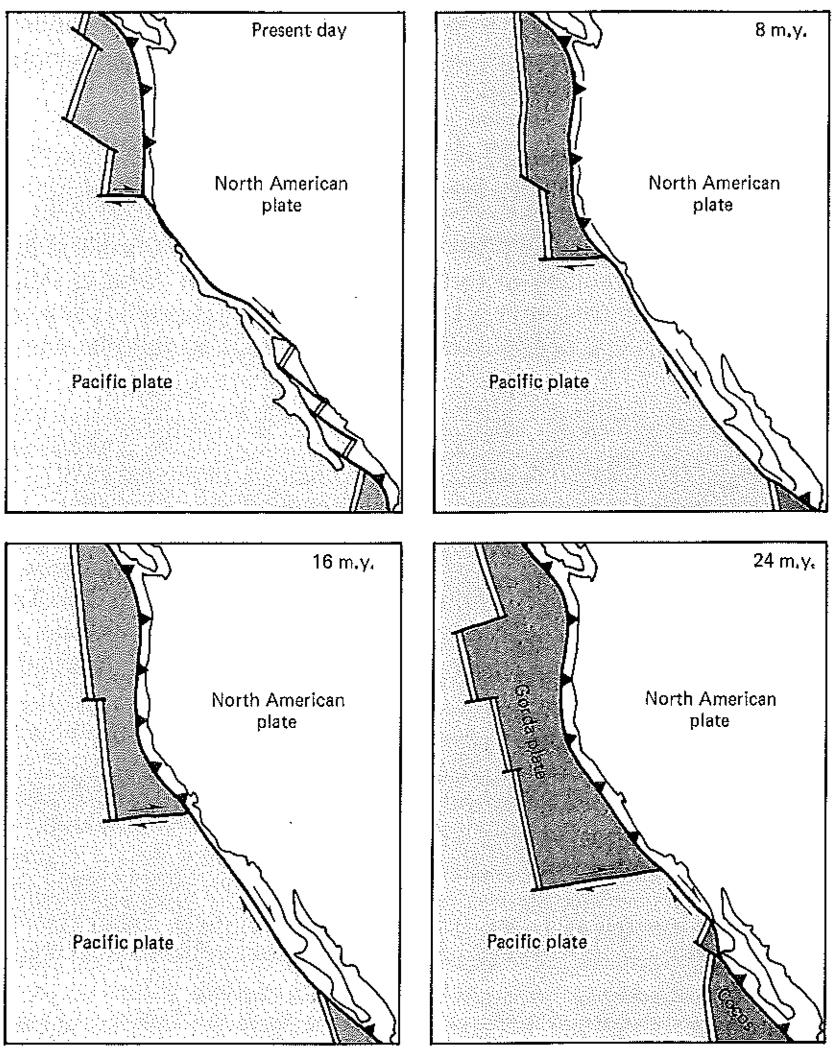

FIGURE 13–36 History of late Cenozoic plate interactions along the Pacific margin of North America. (Compiled from Atwater, 1970, and Engebretson, 1982.)

along the actual continent-ocean boundary. Only fragments of this earlier history of faulting can be assigned to specific faults.

In summary, we can say that reconstructions of plate motions provide a reasonable semiquantitative explanation of the history of late Cenozoic strike-slip that is documented on land from mismatches in geology across individual faults (for example, Fig. 13-30). A second fruitful line of inquiry is to see if the predicted changes from subduction and magmatic-arc tectonics to strike-slip tectonics are actually recorded at the proper places and times in the geologic record. The present areas of Gorda and Cocos subduction have associated magmatic arcs and accretionary wedges. Can we identify a progressive disappearance of these subduction features over the last 30 m.y. associated with the growth of the strike-slip boundary? The answer is qualitatively yes. Marine geophysical studies off the California coast show earlier Cenozoic accretionary-wedge complexes that are now inactive and covered by younger continental-slope sediments that are cut by young strike-slip faults (Fig. 13-37). It is difficult to date precisely the change in tectonics from seismic data available, but it is clear that the change in continental-margin tectonics predicted from plate reconstructions has in fact taken place.

The patterns of magmatic-arc volcanism also show a progressive disappearance in California and Baja California in general agreement with plate reconstructions. However, there is a lot of additional late Cenozoic magmatic activity in western North America, and especially the western United States, that would not have been predicted solely on the basis of plate reconstructions. This additional magmatic activity and related tectonism is the subject of the following sections.

UPLIFT, RIFTING, AND MAGMATISM EAST OF THE PLATE BOUNDARY

The region east of the subducting and strike-slip boundaries of the Cordillera has undergone a remarkable evolution in tectonics beginning about 15 m.y. (Miocene), particularly in the United States. The three most-obvious tectonic phenomena have been regional uplift, normal faulting, and a change in volcanic petrochemistry.

Regional Uplift

The western United States stands regionally high with much of the area above 1.5 km (Fig. 13-38); however, this uplift is anomalous with respect to many other mountain belts, such as the Himalayas, Alps, or Andes, which display a thickening of the crust through strong horizontal compressive deformation. The crust of the western United States is similar in thickness to the cratonic interior (about 40 km) and in some places is much thinner (20–30 km). Furthermore, the

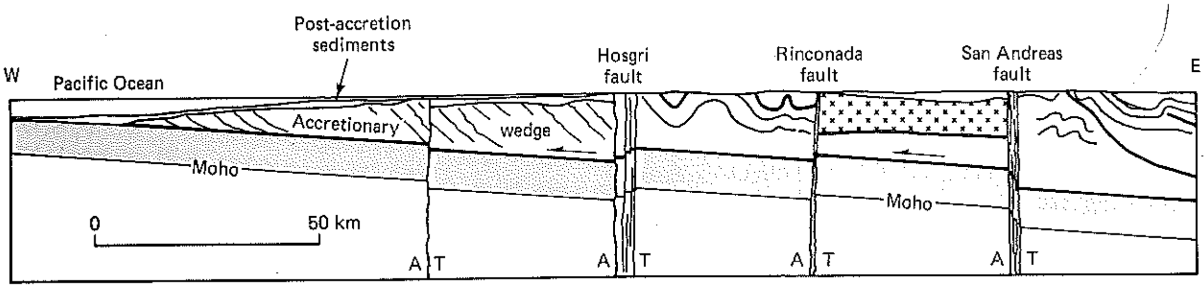

FIGURE 13–37 Cross section of continental margin in central California, showing the earlier Cenozoic accretionary wedge and the later strike-slip faulting. (Based on section of Page and others, 1979.)

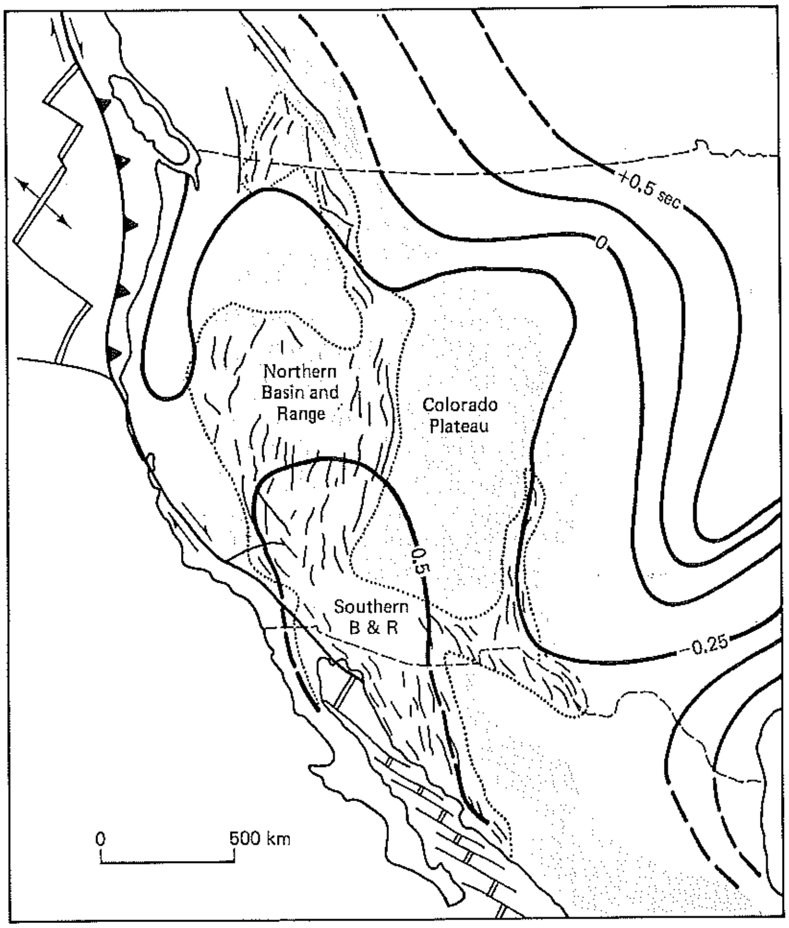

FIGURE 13–38 Map of late Cenozoic plateau uplift and normal faulting. Area of plateau uplift above 1.5 km in screened pattern. *P*-wave travel-time anomalies showing the area of anomalously hot upper mantle in the western United States. (Compiled from Suppe, Powell, and Berry, 1975, and Herrin and Taggart, 1968.)

uplift is epeirogenic in character, similar to the plateau uplifts of the Guiana or Brazilian Highlands or the East African rift system (Chapter 1). The Grand Canyon in the Colorado Plateau (Fig. 13-3) is symptomatic of much of the late Cenozoic uplift; it is a region of typical cratonic geology, but it now stands, largely undeformed, at 1.5 to 2 km above sea level.

We know from the natural experiments in isostasy, such as the unloading of glacial Lake Bonneville near Salt Lake City, that the lithosphere of the western United States is unable to support the present high elevations with its own strength. It must be isostatically supported. Furthermore, early and middle Cenozoic fossil plants from throughout the western United States show that most of the region stood within a few hundred meters of sea level prior to late Cenozoic. Therefore, some major tectonic process must have operated on the lithosphere to produce this regional uplift. There are two principal ways that the lithosphere can be uplifted (Chapter 1): (1) The crust, which is less dense than the aesthenosphere, can be thickened by horizontal compression. This has not happened in the late Cenozoic because the compressive structures and great crustal thicknesses are not observed; (2) The mantle lithosphere, which is more dense than aesthenosphere, can be thinned by heating from below or possibly removed by other processes.

Thinning of the mantle lithosphere can be looked for geophysically. For example, if the upper mantle is abnormally hot, it will transmit seismic waves at lower velocities than normal and cause anomalously late times of arrival of seismic waves from distant earthquakes at seismographic stations. The seismic waves come in nearly vertically from distant earthquakes, and most of the anomaly is caused by differences in upper-mantle velocities directly below the station, relative to the average for continental areas. It is found that the area of high elevations in the western United States shows anomalously slow arrival times (Fig. 13-38). Other types of geophysical measurements that are sensitive to the state of the upper mantle, such as magnetometer arrays, magnetotellurics, and surface-wave propagation, all show an anomalously hot upper mantle and thin mantle lithosphere under the region of high elevation. Therefore, we can conclude that the late Cenozoic uplift of the western United States is an effect of change in temperature and thickness of the mantle lithosphere. The fundamental process is controversial, but the observations show us that we should not look just to the crust to understand late Cenozoic tectonics of the Cordillera.

Normal Faulting

Crustal deformation, in the form of normal faulting, is nevertheless important over about half of the region of late Cenozoic uplift east of the plate boundaries. The bulk of this region is the *Basin and Range* physiographic province (Fig. 13-38) marked by long parallel horsts and grabens. The *northern Basin and Range* is still quite actively faulting and displays extensive internal drainage. It is the site of the 1915 Pleasant Valley earthquake, which produced the scarp shown in Figure 8-6. The *southern Basin and Range* in southwestern Arizona and New Mexico and in Sonora appears to be much less active, displaying more-subdued topography, throughgoing drainage, and fewer earthquakes. Deformation at this latitude is concentrated today in the central rift system of the Gulf of California. Another important locus of Cenozoic extension is the *Rio Grande rift* of New Mexico.

The normal faulting is accompanied by progressive tilting of the fault blocks (Fig. 8-15). For example, the seismic section of the Railroad Valley graben in eastern Nevada (Fig. 13-39) shows progressively greater tilting of deeper reflectors. The tilting of the fault blocks is thought to indicate the flattening of the normal faults (Chapter 8) into a zone of plastic flow in the lower crust.

The amount of horizontal extension is not well known but is generally considered to be substantial, on the order of 100 to 200 km in the northern Basin and Range. The thinnest crust in the western United States, 20 to 30 km thick, is found in the Basin and Range, which is in agreement with large horizontal extension.

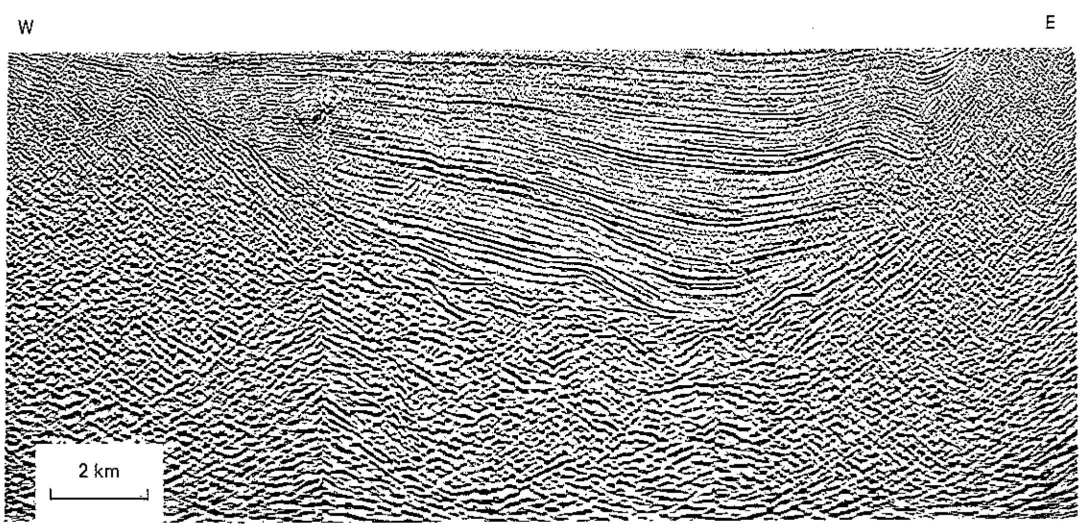

FIGURE 13–39 Seismic section of the Railroad Valley graben, eastern Nevada. (Courtesy John H. Vreeland, Northwest Exploration Company, and Rocky Mountain Association of Geologists.)

FIGURE 13–40 Basaltic lava flows of the Columbia River Basalt, Grande Ronde River, southeastern Washington. (Photograph courtesy of Washington State Department of Commerce and Economic Development.)

Magmatism

The third major aspect of late Cenozoic tectonics east of the San Andreas fault is a major change in the nature of the magmatic activity. The Cenozoic, in general, was a time of voluminous magmatism in the western Cordillera; for example, much of the Mexican Cordillera is covered by Cenozoic tuffs and lavas, which greatly obscures the Paleozoic and Mesozoic history (Fig. 13-2). Sufficient isotopic dating of Cenozoic volcanism has been done in the western United States to assure that the temporal patterns are well defined. Early to middle Cenozoic is marked by voluminous intermediate to silicic welded ash-flow tuffs erupted from great calderas overlying batholithic magma chambers—for example, the Timber Mountain caldera in Nevada (Fig. 8-21). This magmatism is broadly related to subduction along the Pacific margin, generally predating the uplift and Basin-and-Range faulting in the western United States. An important petrologic characteristic of this magmatism is that the bulk of the magma is of intermediate silica content.

About 15 million years ago, broadly coincident with the beginning of uplift and rifting, a change in magmatism began, to a basalt-rhyolite association, characterized by a bimodal distribution of silica content. This change in magmatism was not everywhere simultaneous; for example, today the intermediate magmatism is still active along the subduction-related volcanic chains of the Cascades and Mexico, while the bimodal volcanism is associated with active normal faulting in the Basin and Range and the Rio Grande graben.

A second major development in magmatic activity is the massive outpouring of flood basalts in the Columbia River (Fig. 13-40) composing about 250,000 km^3 of lava, with most of it extruded in just 3 m.y. (16 to 13 m.y.). The Columbia River flood basalts are quite comparable to the other major fields of flood basalts of the world—for example, the Deccan traps of India, the Parana basalts of South America, and the Greenland and Scottish flood basalts. Also broadly associated with the Columbia River flood basalts is the Snake River Plain–Yellowstone volcanic belt (Fig. 13-2), which exhibits some similarity to oceanic hot-spot tracks or aseismic volcanic ridges (Chapter 1). The volcanic belt has been propagating toward the northeast for the last 10 m.y. at about 2.5 cm/year and is approximately fixed with respect to other hot spots, such as Hawaii. Therefore the voluminous magmatism of the Columbia River, Snake River Plain, and Yellowstone may reflect a deeper mantle instability rather than a crustal process. This possibility is supported by observations of seismic travel-time anomalies directly under Yellowstone. A carrot-shaped zone of anomalous mantle exists under Yellowstone, with velocities about 5 percent lower than the already-anomalous surrounding upper mantle of the western United States (Iyer and others, 1981). The anomalous mantle extends to a depth of about 300 km, well below the base of the lithosphere.

EXERCISES

13–1 Compare the contrast the Sonoma and Antler orogenies of the Cordilleran mountain belt with the Taconic orogeny of the Appalachian mountain belt.

13–2 Why does the Upper Carboniferous–Permian clastic sequence thin toward the west, whereas the Cambrian–Upper Precambrian and Lower Carboniferous clastic sequences thin toward the east (Fig. 13-4)?

13–3 Estimate the percent shortening for the entire cross section of the eastern part of the southern Canadian Cordillera in Figure 13-20.

13–4 Compare the nature of the structures developed in the Laramide basement uplifts with the structures of the Cordilleran fold-and-thrust belt. What are the reasons for the similarities and differences?

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