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Tectonics: 50 years after the Revolution

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ABSTRACT

The Plate Tectonic Revolution that transformed Earth science has occurred together with revolutions in imagery and planetary studies. Earth's outer layer (lithosphere: upper mantle and crust) comprises relatively rigid plates ranging in size from near-global to kilometer scale; boundaries can be sharp (a few kilometers wide to diffuse, hundreds of kilometers) and are reflected in earthquake distribution. Divergent, transform fault, and convergent (subduction) margins are present at all scales. Collisions can occur between several crustal types and at subduction zones of varying polarity. Modern plate processes and their geologic products permit inference of Earth's plate tectonic history in times before extant oceanic crust. Ophiolites provide an insight into the products and processes of oceanic crust formation. Ophiolite emplacement involves a tectonic process related to collision of crustal margins with subduction zones. The Earth's mantle comprises, from top to bottom, the lithosphere, asthenosphere, mesosphere, and a *hot boundary layer*. Plume-related magmatism may arise from bulges in the latter, which in turn may alternate with depressions caused by pronounced subduction, leading to assembly of supercontinents. Plate tectonic activity probably occurred on an early Archean, or even Hadean, Earth. Earth-like plate tectonic activity seems not to be present on other terrestrial planets, although strike-slip faulting is present in Mars's Valles Marineris. Possible extensional and compressional tectonics on Venus and an inferred unimodal hypsographic curve for early Earth suggest that Venus may be a modern analogue for a young Earth.

INTRODUCTION

The past 50 years have been ones of revolutionary change in tectonics, marked by the plate tectonic revolution on Earth, and accompanying revolutions in understanding of the tectonics of planets as well as revolutions in travel, communication, computers, remote sensing, and presentation. This article aims to outline what we have learned in the field of tectonics and of the Earth in the past 50 years. To fulfill this task, however, it is necessary to touch briefly on other revolutions, as well.

In his book *The Structure of Scientific Revolutions* (1970), T.S. Kuhn described how scientific revolutions proceed: A science spends the bulk of its history in a "mopping up" phase, wherein a generally accepted unifying model, or paradigm, forms the basis of most research projects. New information fills in the "nooks and crannies" of that reigning paradigm. Information that

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doesn't fit with the reigning paradigm is discarded or put aside. At some point in development of a science, so much information accumulates that doesn't fit the reigning model that the science undergoes a crisis. Many workers recognize that the old model does not work, but as yet there is no new idea as to what might be better. Eventually, someone comes along with a new idea to incorporate all the new contradictory information with the old model to develop a new one that has greater explanatory power.

Such it has been with tectonics. If one goes back 53 years, to the fall of 1959, the reigning paradigm in tectonics was the geosynclinal hypothesis. According to this model, continents and ocean basins were fixed, and oceans were as old as continents. From time to time, long furrows or geosynclines developed with volcanic-free marginal zones, or miogeosynclines of thick sedimentary rocks, and a central volcanic-rich eugeosyncline of deformed intrusive and extrusive igneous and metamorphic rocks. Deformation was oldest in the central zone and moved outward in all directions, emplacing eugeosynclinal rocks over miogeosynclinal rocks and continental shelves, respectively.

Mountain (orogenic) belts formed from such processes. Tectonics was primarily a regional descriptive exercise, with little or no constraint. One prominent geology department (Caltech) did not even have a course in regional structural geology, let alone tectonics. Alfred Wegener's (1912, 1922) idea of continental drift (reportedly including a reference to the Mid-Atlantic Ridge as a zone of ocean floor spreading: Wegener (1912, *in* Jacoby, 2012) was widely, but not completely, disregarded in North America, although less so in Europe, and was widely accepted by South African geologists (e.g., Oreskes, 1999).

In the late 1930s, evidence for thin oceanic crust and earthquake concentrations along the mid-Atlantic ridge had led people in both Europe and North America to reconsider Wegener's hypothesis (e.g., Heck, 1938; Field, 1938; Hess, 1938; Stille, 1939). The International Geological Congress scheduled for New York in September 1939 had as a main focus the reconsideration of Wegener's hypothesis. The German invasion of Poland and the start of World War II the week before aborted the Congress, and reconsideration of Wegener had to wait.

During and after World War II, enormous developments in acoustic ocean bottom sounding, seaborne and airborne magnetometers, gravity measurements on surface ships, and oceanic seismic exploration led to a flood of new information on the oceans (e.g., Menard, 1986; Hess, 1946, 1948; Hess and Maxwell, 1953). By 1959 it had become clear that the geosyncline–permanent continent–ancient-ocean-basin hypothesis was in crisis. Polar wander paths, developed by P.M.S. Blackett, S.K. Runcorn, and others in the 1950s (e.g., Runcorn, 1962) indicated that the position of the Earth's magnetic pole had changed with respect to the continents. Seismic work confirmed that ocean crust was thin and stood isostatically 5 km or so below continents, a discovery that had been called "the most momentous since the 2nd world war" (Hess, 1954; see also Menard, 1986, and Moores, 2003). Mid-ocean ridges were global in extent (Heezen, 1960) and had high heat flow and concentrations of earthquakes (von Herzen, 1959; Bullard et al., 1956). Sediments in the oceans were thin (Menard, 1956, 1986). Earthquakes around the Pacific were aligned along zones dipping beneath island arcs (e.g., Benioff, 1955; Hess, 1948). Geology was ripe for a revolution.

The spark arguably came from Harry Hess of Princeton University. The event that possibly set it off was a talk by S.W. Carey in the fall of 1959. At that time Hess had been lecturing to first-year geology graduate students, including one of us (EMM), in a course entitled "Advanced General Geology," expressing skepticism about polar wander paths, emphasizing problems of paleomagnetism, discussing how "Alpine serpentines are probably intruded during the first great deformation of a given mountain system" (Hess, 1955, p. 395), and outlining what was known about the oceans—thin crust, high heat flow over ridges, trenches, etc., and reiterating, "So the problem remains unsolved. Some vital piece of evidence is still missing" (Hess, 1955, p. 402). In December 1959, Carey lectured in the Princeton University Geology Department on his ideas on continental drift (Carey, 1958), involving, among other items, his reconstruction of the fit of continents about the Atlantic Ocean, using the 100- and 1000-fathom contour lines—i.e., the continental edge (rather than the edge of the shelf or the coastline, as Wegener had done), and plotting reconstructions on a 30-in-diameter spherical table; thus, in those pre-computer years, dispensing with the problem of map projection distortions, and calling on Earth expansion to explain the opening of the Atlantic Ocean.

In the middle of Carey's talk, Hess bolted out of his seat and spent the rest of the 3-h lecture pacing back and forth, chain-smoking, in the room (it is remarkable how much secondary smoke one dealt with in those days). Subsequently there was no more discussion of problems of paleomagnetism in Advanced General Geology. About six weeks later, Hess was circulating a manuscript entitled "Evolution of Ocean Basins" in which he proposed spreading of the ocean floor from the ridges outward (Fig. 1) (the paper appeared in 1962 with the title "History of Ocean Basins" (Hess, 1962). As it was then known that the mineral serpentine was stable at temperatures below 500 °C, Hess

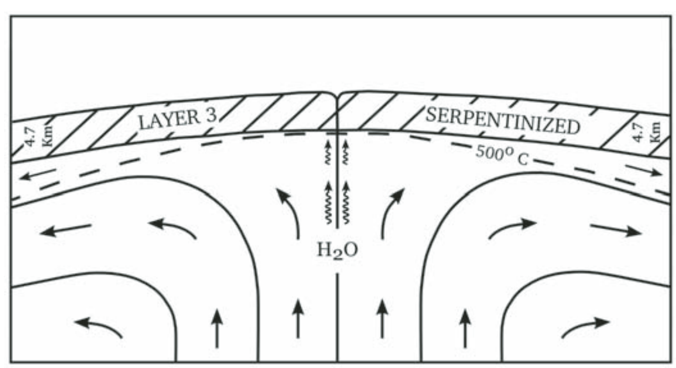


Figure 1. Diagram of ocean floor spreading, first outlined by H.H. Hess in 1960 (published as Hess, 1962). Formation of ocean crust produced by serpentinization of a spreading Earth's mantle. Hess suggested that the motion resulted from mantle convection (from Hess, 1962, Fig. 6).

reasoned that the ocean crust–mantle boundary was a serpentinization front, although he later accepted the volcanic nature of the upper oceanic crust (Hess, 1964). Hess did not deal in detail with return of the material back into the ocean. However, he did invoke convection and noted that "mantle convection is considered a radical hypothesis not widely accepted by geologists and geophysicists. If it were accepted, a rather reasonable story could be constructed to describe the evolution of ocean basins and the waters within them. Whole realms of previous unrelated facts fall into a regular pattern, which suggests that close approach to satisfactory theory is being attained" (Hess, 1962, p. 607).

With first, the circulation, and then the publication (1962) of Hess's article, events started to happen at a quickening pace. Dietz (1961) coined the term *sea floor spreading,* Vine and Matthews (1963) related changes in polarity of the Earth's magnetic field to symmetrical magnetic anomalies about the Carlsberg (Indian Ocean) ridge; Vine (1966) subsequently extended his analysis to all the oceans, and Wilson (1965) elaborated on ideas expressed informally by Fred Vine, defining *transform faults* on mid-oceanic ridges. Bullard et al. (1965) published a computerized fit of continents around the Atlantic Ocean (see also Everett and Smith, 2008). Dickinson and Hatherton (1967) summarized the relationship between andesites and circum-Pacific volcanoes. Several papers in 1968 (e.g., Isacks et al., 1968; Heirtzler et al., 1968; Morgan, 1968) developed the general outlines of the "New Global Tectonics." Isacks et al. (1968) documented the existence of dense, high-velocity mantle beneath oceans, and dipping beneath island arcs, and concentration of earthquakes near ridges, trenches, and transform faults. Morgan (1968) outlined the geometric features of plates (originally called *blocks*), and Heirtzler et al. (1968) presented a global correlation of magnetic anomalies.

In 1969, William R. Dickinson organized a Penrose Conference on "The Meaning of the New Global Tectonics for Geology" (Dickinson, 1970a). At the meeting, talks on issues such as ophiolites, the Appalachian-Caledonide belt, the relations between island arc volcanism, seismicity, "underflow" of mantle along the U.S. Pacific margin, and actualistic examples of "geosynclines" made it clear that the "New Global Tectonics" indeed extended into geologic time at least as far back as the Paleozoic Era (Dickinson, 1971). An issue that arose was what to call convergent margins. Debate raged over the merits of "consumption zones" or "subduction zones (subduction as a translation of the term *Verschluchung*—"swallowing"—used by Ampferer and Hammer to explain the Alps in 1914; cf. Roeder, 1973). The issue was settled by Conference elder statesman John C. Crowell, who said that he "would rather be subducted than consumed" (cf. White et al., 1970).

Developments moved quickly after that. In the next few years, a number of papers emerged (many had been presented at the Asilomar conference) that connected the newly recognized plate tectonic concepts to the rock and evolutionary records. Leg 3 of the Deep Sea Drilling Project (Maxwell et al., 1970) in the south Atlantic ocean established an age progression of oceanic crust away from the mid-oceanic ridge, in line with Hess's (1962) and Vine and Matthews'(1963) hypotheses. Examples of new views of the Earth included studies of island arcs and marginal basins of the western Pacific (Karig, 1971, 1972, 1974), the plate tectonic origin of the western United States (e.g., Hamilton, 1969a; Burchfiel and Davis, 1972), and the interpretation of blueschists of the Alps, California, and Japan as metamorphism in a subduction zone (e.g., Ernst, 1970, 1971, 1973; Oxburgh and Turcotte, 1971). Also discussed were the nature of mélanges and their relation to subduction (Hsü, 1971a); the nature and origin of subduction-related volcanic and plutonic rocks (e.g., Dickinson, 1970b; Hamilton, 1969b; Marsh and Carmichael, 1974), the origin and emplacement of ophiolites as fragments of ocean crust and mantle formed at oceanic spreading centers (e.g., Moores and Vine, 1971; Moores, 1970; Coleman, 1971; Dewey and Bird, 1971). Also added to the mix were discussions of plate tectonics and development of the Alpine-Himalayan belt (e.g., Smith, 1971; Hsü, 1971b; Dewey et al., 1973) as well as the Appalachian-Caledonides (Dewey and Bird, 1970) and the Andes (Hamilton, 1969b). Other topics were relating modern and ancient sedimentary sequences to the plate tectonic environment of their formation (Dickinson, 1971); the driving mechanisms, causes, and other aspects of plate tectonics (McKenzie, 1969); plate tectonics and mantle convection (e.g., Turcotte and Oxburgh, 1972); plate tectonics and ore deposits (Sillitoe, 1972; Sawkins, 1972; Mitchell and Bell, 1973) and the relationship between patterns of continental drift and evolution (e.g., Valentine and Moores, 1970, 1972). Hamilton (1969a) also pointed out that accretionary prisms include sedimentary material derived from the adjacent island arc or continent, as well as from the deep oceans. Dickinson (1970b) established the relationship between andesites and granitic intrusive complexes and subduction zones. Morgan (1971, 1972) developed evidence for rising plumes of hot solid mantle originating from the lower mantle or core-mantle boundary. These revisions extended the impact of plate tectonics to all aspects of geology, and through much of Phanerozoic time. It was an exciting time.

The Plate Tectonic Revolution did not receive universal overnight acceptance. Some field geologists, both in North America and in Europe, imbued with a sense of the value of field exposures, especially in the Alps and western North America, and wedded to an inductive, versus a model-deductive, approach to study of the Earth, resisted plate tectonics for several years, (e.g., Meyerhoff and Meyerhoff, 1970; Gilluly, 1971, 1973; cf. Schaer, 2011). For example, a famous geologist at a prestigious Alpine geology department denounced an early manuscript of a plate tectonic analysis of the Alps in early 1973.

Plate tectonics are driven by ocean-derived investigations and processes, as well as by both geophysical and geochemical, as well as geological, evidence. One did not need to have years of experience in the Alps or western North America to contribute to an understanding of these regions. Indeed, Trümpy (2001) suggested the small sizes of the Swiss and Austrian navies as a possible reason for the lack of Alpine geologists in the development of plate tectonics. Many senior geologists ultimately accepted the theory (e.g., Trümpy, 2001); others did so fairly rapidly (e.g., Bernoulli and Jenkyns, 1974; Laubscher, 1971). Eventually, opposition to plate tectonics receded as time passed and more evidence was accumulated. The role of model-deductive versus inductive reasoning in the geosciences still is a subject of disagreement, however.

CURRENT EARTH VIEW

Since the Plate Tectonic Revolution, one of the great scientific revolutions of all time, we have learned much new information about the Earth and terrestrial planets. This article will highlight some salient features of what we have learned. In the space available, it is not possible to cover all topics thoroughly; indeed, several books have been written on the subject (e.g., Condie, 1997; Cox, 1973; Cox and Hart, 1986; Moores and Twiss, 1995; Twiss and Moores, 2007, Kearey et al., 2009). Acquisition of much of the information we now have on the Earth postdates the Plate Tectonic Revolution. The Deep Sea Drilling Project (DSDP) and its successors, the Ocean Drilling Program (ODP) and the Integrated Ocean Drilling Program (IODP), have added an enormous amount to our understanding of the oceans. Since 1970 seismic reflection studies (e.g., the Consortium for Continental Reflection Profiling, COCORP); Oliver et al., 1976) also have added hugely to our knowledge of the structure of the continents. Many geophysical studies provided new insight of Earth's interior. Revolutions in information handling and geological, geophysical, and geochemical information gathering have contributed to this surge in knowledge as well. The global positioning system (GPS) has recently enabled determination of the motion of tectonic features in real time. Deployment of seismic arrays using PASSCAL instruments (Aster et al., 2005) has made it possible to image the interior of the Earth at unprecedented resolution (Dziewonski, et al., 2010; Lay and Garnero, 2011); dense arrays such as U.S. Array, combined with techniques such as ambient noise tomography, increase the resolution of tomographic maps (Yao et al., 2006; Lin et al., 2010). Revolutions in radiometric dating, particularly involving new instruments and isotopes of potassium, uranium, thorium, and other elements, have greatly increased our knowledge of the age of rocks and structures. Revolutions in information handling, and in geological, geophysical, and geochemical information gathering, have contributed to this surge in knowledge as well. Space geodetic techniques such as very long baseline interferometry (VLBI) (Davis et al., 1985), and more recently the GPS (Hager et al., 1999) and interferometric synthetic aperture radar (InSAR) have enabled determination of the kinematic motion of tectonic features (e.g., McClusky et al., 2000) and global plate motions (Argus et al., 2010) in real time.

MAJOR FEATURES OF THE EARTH

The Earth contains three principal zones—the crust, mantle, and core (Fig. 2). The crust is principally divided into continental and oceanic crust.

Continental crust averages ~35 km thick, being thinner at continental margins and thicker underneath orogenic belts, and it is as old as ca. 4.0 Ga (Bowring and Williams, 1999), with some minerals being even older (zircons from the Jack Hills, Australia, are 4.4 Ga; Wilde et al., 2001; Harrison, 2009). Continents exhibit margins, interior platforms, Precambrian shields, and orogenic belts. Continents in general appear to be a composite of old pieces of crust and younger orogenic belts (e.g., Hoffman, 1988). Precambrian shields extend beneath platforms and are involved in several younger orogenic belts. The interior platforms are mostly horizontal sedimentary rocks on top of these Precambrian shield regions. The ages of consolidation of continents differ from place to place. Major periods of consolidation include two periods of supercontinent (collection of all continents into a single landmass) formation (Rodinia, ca. 900 Ma; Pangaea, ca. 250 Ma) and possibly some older ones as well (e.g., Nuna or Columbia, ca. 1800–1900 Ma until ca. 1300 Ma; e.g., Rogers and Santosh, 2002; Evans and Mitchell, 2011). Major peaks of continental crust ages occur at 2.7 Ga, 1.9 Ga, 1.2 Ga, and less than 0.4 Ga (Harrison, 2009).

The oceanic crust is strikingly uniform in thickness, averaging ~7 km thick; but it is thicker under oceanic island arcs, along hotspot tracks, and beneath oceanic plateaus. The oceanic crust in the modern oceans is all less than 200 m.y. old; the oldest oceanic crust is ca. 180 Ma in the western Pacific plate, just east of the Marianas (National Geophysics Data Center, 1996).

Figure 3 illustrates a recent view of the Atlantic, Pacific, and Indian Ocean basins, based on gravity satellite measurements (Matthews et al., 2011; Bird, 2003). Present are the mid-oceanic ridges, transform faults, oceanic plateaus, aseismic ridges (hotspot traces), and fracture zones—the scars of formerly active transform faults.

The oceanic crust seismic structure was originally thought to consist of three layers: layer 1 of sediments; layer 2, generally thought to represent extrusive lavas; and layer 3, the main oceanic layer, formed of dikes, gabbros, and serpentinized peridotite. Current thinking, however, is that the oceanic crust is more complex, and the seismic layers result more from porosity and alteration effects than from igneous lithologies (e.g., Karson, 1998).

The mantle's topmost layer is the lithosphere, which is negligible in thickness at oceanic spreading centers but thickens to an average of ~80 km beneath oceans and up to several hundred kilometers underneath continents. Beneath the lithosphere is the asthenosphere, which is thought to represent a region closer to its melting temperature; thus it is of lower S-wave velocity.

The asthenosphere is underlain in turn by the mesosphere, from 660 km to the core-mantle boundary, 2900 km deep.

In the mesosphere the principal mantle mineral, olivine, converts to the more dense mineral Mg-perovskite. Near the core-mantle boundary, seismic velocities change again in the region known as the D″ layer (e.g., Bullen, 1950).Two large, lower mantle regions display low S-wave velocities and high density (Romanowicz, 1998), centered approximately on the equator. As part of mantle convection, plumes or hotspots are thought to rise

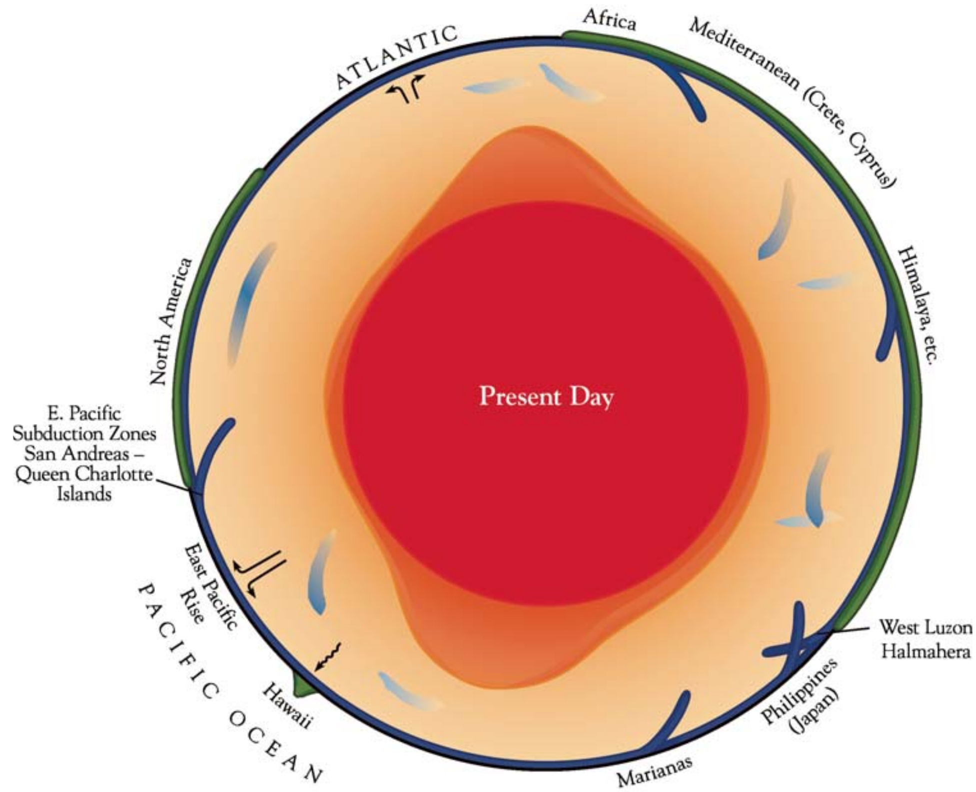


Figure 2. Schematic cross section of present-day Earth. Irregular zig-zag profile (or sector) through the Earth shows continental crust (green), oceanic crust (black), lithospheric plates (blue), lower mantle (yellow), hot abyssal layer (light orange), and core (darker orange). Light-blue streaks are schematic representations of remnant slabs in lower mantle. Modified after Kellogg et al. (1999). Not to scale.

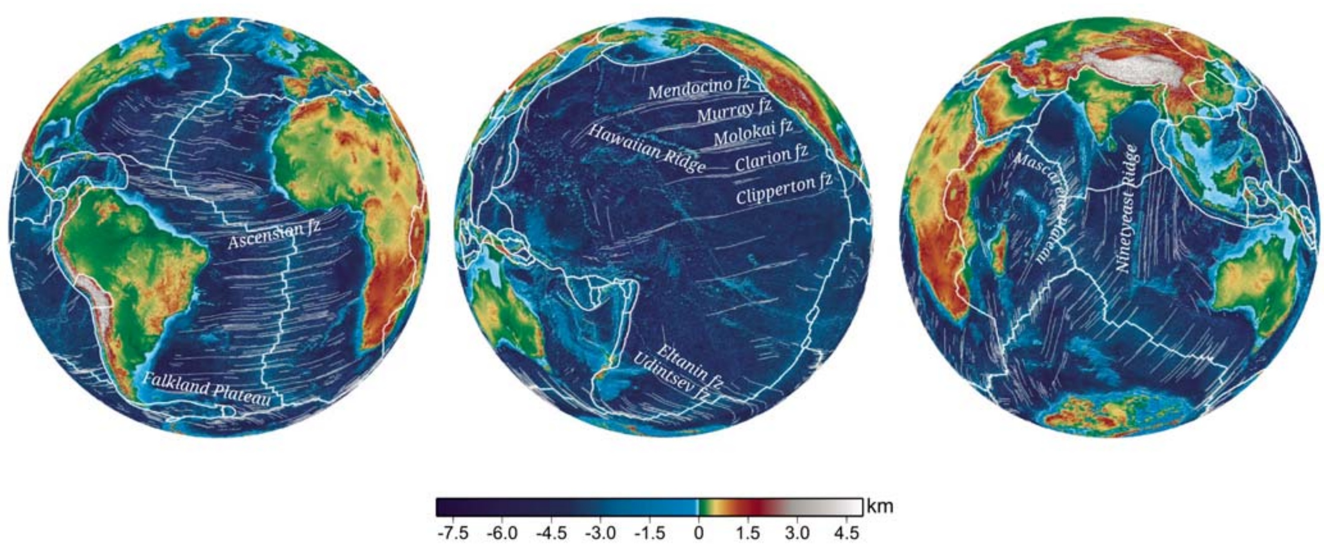
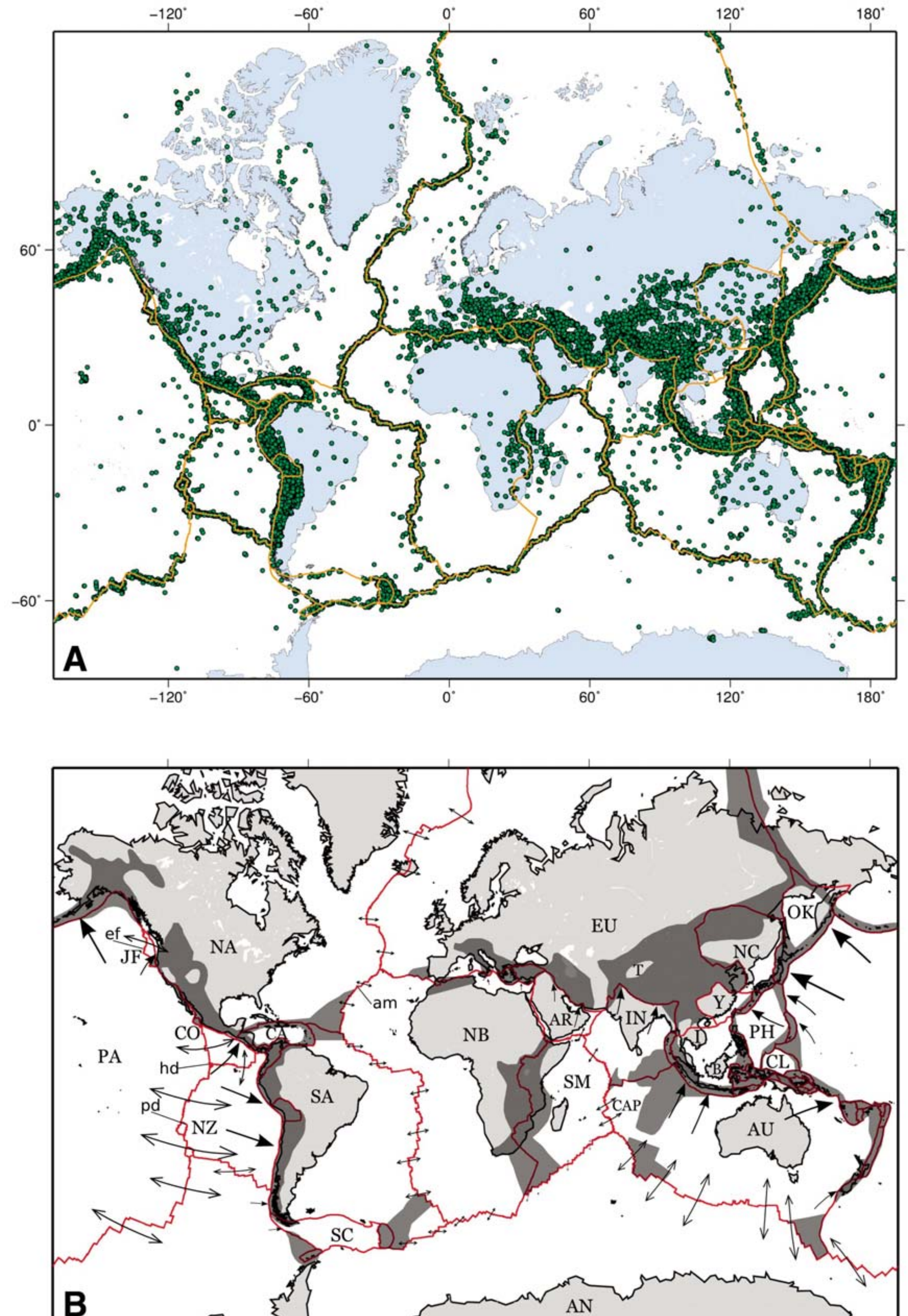


Figure 3. Illustrations of oceanic basins, showing Atlantic, Pacific, and Indian Oceans. White lines are main plate boundaries. Light-blue linear features are aseismic ridges, similar to the Hawaiian Ridge. Selected fracture zones and transform faults are named. Modified after Matthews et al. (2011) and Bird (2003).



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from unstable hot boundary layers (possibly at the core-mantle boundary or from the top of an "abyssal hot layer" (Kellogg et al., 1999) to form aseismic ridges, such as the Hawaii-Emperor chain or the Greenland-Iceland-Scotland volcanic ridges (see Figs. 2, 3). Seismic tomography has revealed a great deal of the structure of the lower mantle, a region of high velocity beneath Asia, and regions of low velocity beneath Africa and the SW Pacific. Burke (2011) suggested that these regions are long-lived, stationary, and approximately antipodal, and that plumes may arise from around these features.

The core consists of an outer liquid portion of Fe-Ni alloy. Currents within this part give rise to the Earth's magnetic field. The inner core is solid Fe-Ni and displays its seismically fast direction oriented parallel to the Earth's rotation axis. The inner core may rotate relative to the Earth as a whole (Su et al., 1996). Core dynamics are largely outside the scope of this paper and so will not be discussed further.

PLATES

Plates are the regions of the Earth's crust separated by plate margins, exhibiting abundant earthquakes (Figs. 4A, 4B). Plate margins are fairly well defined in the oceans, but they are more complex within or next to continents (Fig. 4B). There are seven major plates and many minor plates, ranging in size, depending on one's definition of a plate, from thousands to a few tens of kilometers in extent. New plate material forms at mid-ocean ridges, in backarc basins, within arcs, or in forearc regions; plates move past each other at transform faults, and one plate slides beneath another and into the mantle at convergent margins or subduction zones.

Plates are, of course, principally defined by the distribution of earthquakes (Fig. 4A). A key component of the development of plate tectonics was the recognition of "first-motion" studies from seismology (e.g., Wilson, 1965; Isacks, et al., 1968). These so-called *focal mechanisms* have given rise to so-called *beach-ball* diagrams—stereographic or equal-area projections with quadrants indicating compression (P) and rarefaction (T) of the first motions recorded by seismographs at varying distances from the earthquake. These diagrams nearly universally are considered to provide P and T axes, or major and minor compressive principal stress directions. Twiss and Unruh (1998) suggested that these diagrams reflect the strain-rate field during an earthquake, rather than the stress field.

Several forces affect plate motion. These forces include the pull of the dense slab, and push at the ridges, so-called *trench rollback,* wherein the downgoing-slab bend moves away from the subduction zone as motion proceeds (Gvirtzman and Nur, 2001; Heuret and Lallemand, 2005); and forces related to plumes (Morgan, 1971, 1972; Hill et al., 1992; Cande and Stegman, 2011). Principal resisting forces are interplate friction, bending of the slab, mantle drag, and the transition from spinel or garnet to perovskite (Van Huenen and Moyen, 2012).

McKenzie and Parker (1967) and Morgan (1968) showed that linear motion on the Earth's spherical surface could be explained by angular rotation of one plate relative to another about a pole of rotation. For the oceans, they used the directions of transform faults as lines parallel to relative plate motion (Fig. 5), because for energy conservation reasons, plate motion is nearly invariably parallel to transform faults and perpendicular to the trend of ridges.

Three plates meet in triple junctions. There are 27 theoretically possible triple junction configurations, but only triple junctions involving three ridges are stable in all configurations (Fig. 4B: McKenzie and Morgan, 1969, fig. 3.) See also Cox (1973) and Moores and Twiss (1995, fig. 4.2.4). Dewey (1975) showed that in a collection of three plates, the pole of rotation of at least two plates must move instantaneously with respect to the plates themselves. Thus on Earth, the poles of rotation of several two-plate systems move with respect to the plates themselves. This fact results in shifts of relative motion, migration of the poles of rotation relative to plate boundary orientations, and thus in changes in the trend of features produced by plate motion.

Not long after the formulation of plate tectonics, instantaneous numerical models of global plate motion appeared (e.g., Minster et al., 1974; Minster and Jordan, 1978; DeMets et al., 1990). These models have facilitated analysis of individual plate motions, and several web-based plate motion tools have become available.

In a major refinement of the plate tectonic view of the Earth, McKenzie (1970) and Gordon (1998) pointed out that the original plate tectonic interpretation is an *approximation* that does not fit all areas. Indeed, close-spaced earthquakes are present only on some oceanic plate boundaries. Other oceanic plate boundaries (e.g., the SE Indian Ocean) and boundaries involving continents do not have such a sharp definition. Figures 4A and 4B show the distribution of earthquakes and plates with idealized narrow boundaries (mostly ridges and oceanic transform faults) as well as diffuse plate boundaries. Most diffuse boundaries involve continents, but some are present in oceans, as well, for example, in the Indian Ocean and in the central and southern Atlantic Ocean. The recent Indian Ocean

Figure 4. Mercator projection of Earth, showing earthquakes, major plates, and plate boundaries. (A) Major plate boundaries (yellow) and epicenters of M4 or larger earthquakes in the past 10 years (in green). After Bird (2003), and the ANSS Earthquake Catalog (2012). (B) Major plates and zones of distributed activity. Red lines: plate boundaries. Lines with divergent open arrows depict spreading centers (ridges); closed arrows, subduction zones with length of arrow proportional to convergence rate; arrow on downgoing plate. Earthquakes are relatively narrowly distributed on ridge and transform boundaries. Shaded areas are regions of diffuse earthquake distribution, mostly within continents and along convergent margins. Abbreviations: am—Atlantis massif; AN—Antarctic; AR—Arabia; AU—Australia; B—Borneo; CA—Caribbean; CAP—Capricorn; CL—Carolines; CO—Cocos; ef—Blanco Transform escarpment; EU—Eurasian; hd—Hess Deep; I—Indochina; IN—Indian; JF—Juan de Fuca; NA—North American; NB—Nubian; NC—North China; NZ—Nazca; OK—Okhotsk; PA—Pacific; pd—Pito Deep; PH—Philippine; SA—South America; SC—Scotia; SM—Somali. Note triple junctions. Modified after Gordon (1998, fig. 2, p. 621) and Bird (2003).

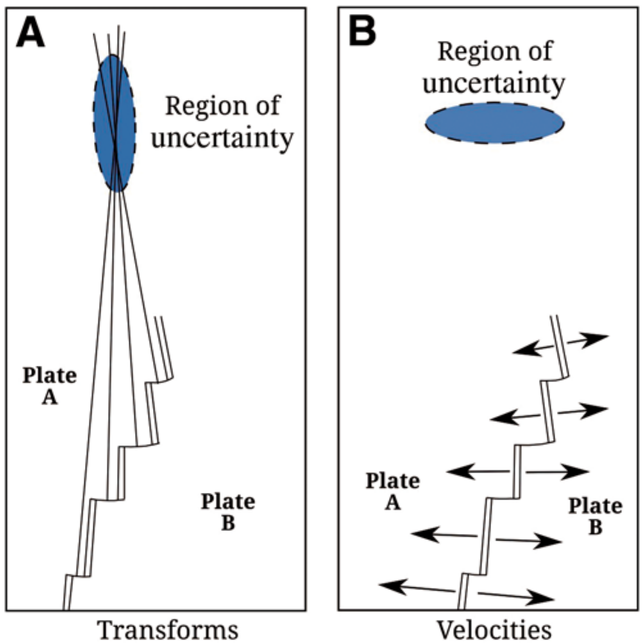


Figure 5. Schematic diagrams, showing errors in determination of Euler poles. Shaded ellipse represents the Euler pole between plates A and B. (A) Region of uncertainty in determination from great circles normal to transform faults. (B) Region of uncertainty from distribution of spreading velocities along a divergent margin (redrawn after Cox and Hart, 1986; see also Moores and Twiss, 1995, fig. 4-5, p. 54).

earthquakes may represent development of a new plate boundary between the Australian and Indian plates (Yue et al., 2012).

Diffuse boundaries cover ~15% of the Earth's surface. In such regions the poles of rotation between the plates are near or within the boundaries. Models for deformation in regions of diffuse plate boundaries include accommodation either by small blocks, or continua, or a combination of these features (e.g., Thatcher, 2009); see Fig. 6.

DIVERGENT BOUNDARIES

Divergent boundaries are where plates move apart from one another. Divergent boundaries commonly begin with updoming and incipient rifting of continents and progress to mature ocean basins. Continental rifting is the subject of a separate chapter in this volume (Ebinger et al., 2013), so it will be mentioned only briefly here. A model for progressive development proceeds from updoming, as in Africa, through development of a narrow ocean region (e.g., Red Sea, Gulf of Aden), then finally to open ocean conditions. A ridge-ridge-ridge (RRR) triple junction ideally develops in such a situation, or one branch may fail, leading to development of a sharp curve in the continental margin and an oblique ridge-transform boundary, as in the central Atlantic.

Four active or recently active intracontinental rift systems on Earth are noteworthy (Fig. 7). The most famous of these are the East African Rift and the North American Basin and Range Province. Other rift systems worthy of mention are the Weddell Sea–Ross Sea region of Antarctica, active during Tertiary time, and the Gamburtsev–SW Indian rift system, perhaps active from Permian-Cretaceous time (Ferraccioli et al., 2011). This latter is interesting, especially because the Gamburtsev Mountains lie beneath at least 1 km of ice; they have been called "the least known place on Earth" (Ferraccioli et al., 2011).

Rifting in many cases seems to be of asymmetrical nature, with packets of subparallel normal faults producing half grabens (Fig. 8). Half-graben stacks alternate with each other along strike in a rift system. Both the East African Rift and the Basin and Range Province show such features. Most continental and oceanic rifts are asymmetrical, with one side representing a lower plate. Rifts are volcanic rich if abundant volcanism accompanies them, or volcanic poor in cases of little or no volcanism. Margins in the present ocean seem to alternate between volcanic-rich and volcanic-poor types.

Core complexes are regions in both continents and oceans where rifting, updoming, and erosion has progressed to the stage where a subhorizontal or domed detachment fault overlies ductilely deformed basement rocks, or in the case of oceanic regions, of lower mafic crust and mantle rocks; the fault underlies brittlely deformed upper crustal rocks (e.g., Tucholke et al., 1998). The right side of Figure 8 schematically represents a possible section beneath the detachment fault that dips to the left.

Oceanic ridges themselves display variations in ridge structure, depending upon the rate of spreading and the abundance of magma supply (Fig. 9). Slow-spreading ridges tend

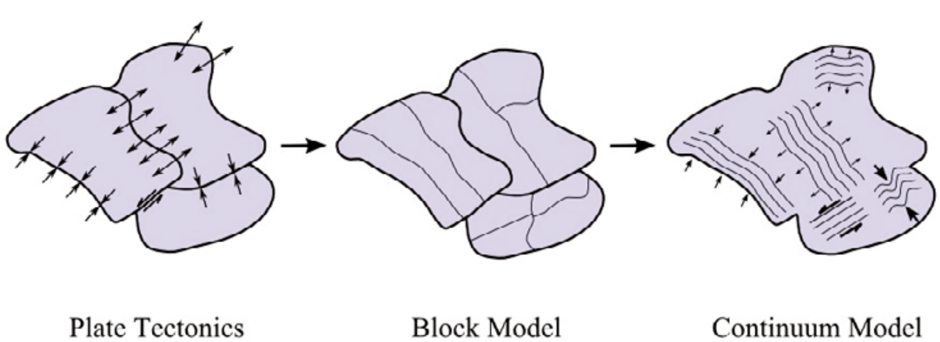


Figure 6. Model for transition from a global plate-tectonic model to block models to continuum models. Simple plate boundaries at global scale may have blocks along the major boundaries at a more regional or local scale, giving way to continuous or continuum behavior between plates. Redrawn after Thatcher (2009, fig. 3).

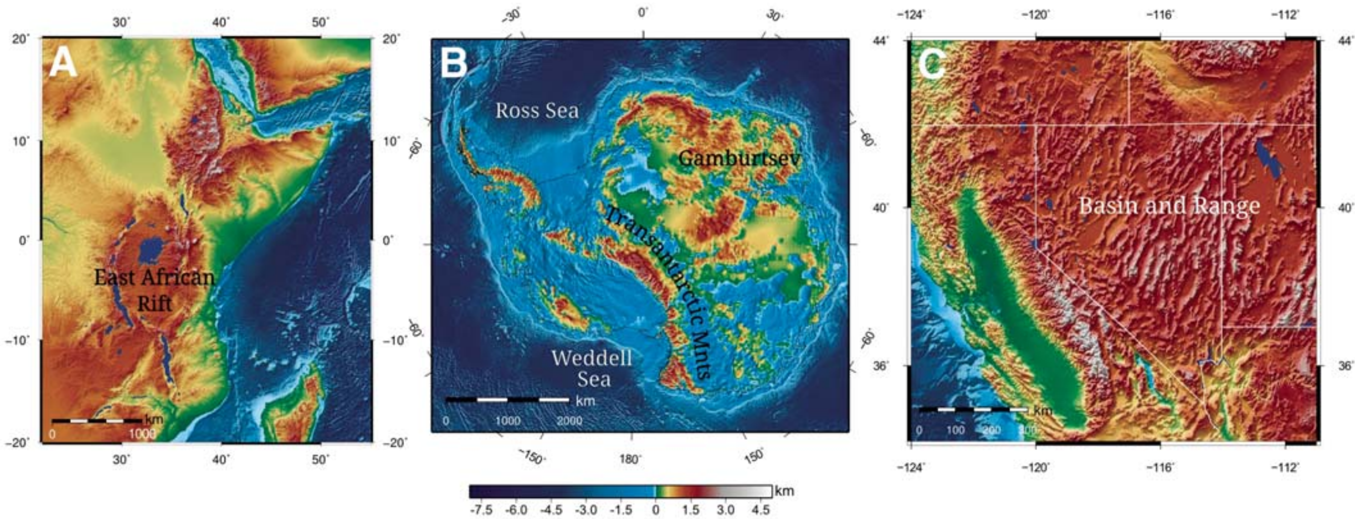


Figure 7. Digital elevation models of the world's major continental rifts, illustrating large elevation variations. (A) East African rift system, showing outline of Western and Eastern Rifts and relations to the Red Sea and the Gulf of Aden rift systems (modified after Burke, 1980). (B) Antarctic rifts: Ross Sea–Weddell Sea rift between West Antarctica (from Behrendt, 1991; Studinger et al., 2006, and www.ldeo.columbia.edu/~mstuding/tam_map_large.html) and the Gamburtsev Mountain rift system (after Ferraciolli et al., 2011). (C) Basin and Range Province, western North America, modified after Stewart (1978). Note large elevation contrasts and differences in scale. Elevation models after Amante and Eakins (2009).

to have a pronounced axial valley with isolated volcanoes and exposures of serpentinized peridotite. Detachment faults are present along many slow-spreading ridges (e.g., Karson, 1998; Tucholke, and Lin, 1994); see Figure 8. Fast-spreading ridge surfaces are smooth, with little or no normal faulting. Intermediate-spreading ridges show topography intermediate between the two end members.

For some very slow spreading-rate, magma-starved ridges, there is no magmatic crust at all, and the oceanic crust is made up of serpentinized peridotite, as originally proposed by Hess, (1962; see Fig. 2). In many cases the serpentinite appears to be the lower plate of a detachment fault, with pronounced "megamullions," ridges oriented parallel to the spreading direction (Tucholke et al., 1998, Blackman et al., 2002).

All ridges display deflections at transform fault intersections, as discussed below. Magma intrusion-extrusion tends to overlap with tectonic rifting, resulting in a varying amount of mixture of magmatic and faulting features.

Direct observations of oceanic crust (Karson, 2002) at three separate windows, the Hess Deep Rift (HD), a hole through fast-spreading crust that formed the Galapagos Rift; the Blanco Transform escarpment, an exposure of intermediate-rate spreading along the Juan de Fuca Rift; and the Pito Deep (Hayman and Karson, 2009), an exposure in the small Easter microplate

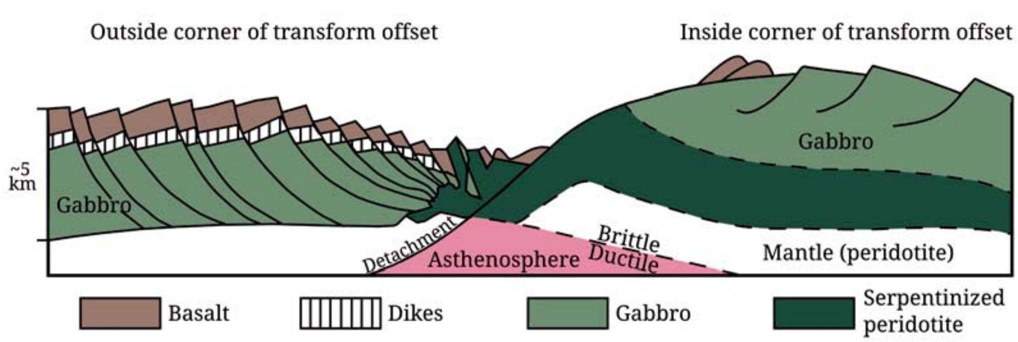


Figure 8. Example of asymmetric spreading along a slow-spreading ridge. Schematic cross section of a faulted ridge near a transform fault zone, with a complete crustal sequence on the left, flanked by an eroded sequence on the right. Sequence on the left (outside corner) represents a corner near an inactive fracture zone; the sequence on the right represents a section near an inside corner, bordering the active transform fault. Zones of serpentinization of mantle peridotite are shown schematically. Redrawn after Lagabrielle et al. (1998) and Twiss and Moores (2007, fig. 19.16, p. 594).

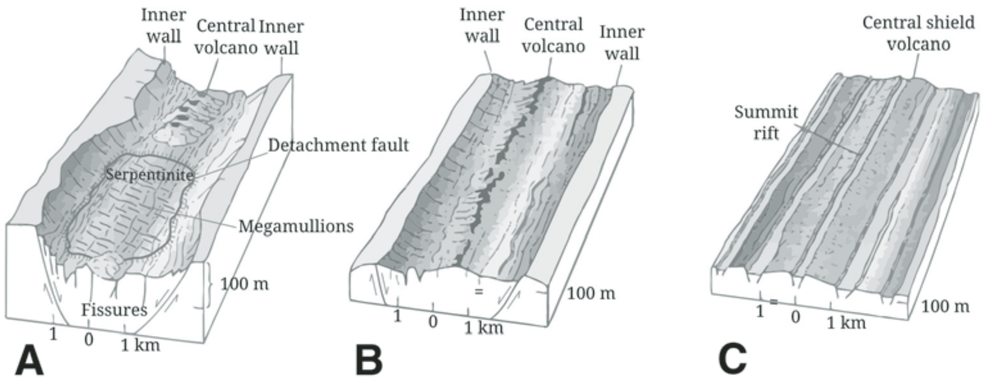


Figure 9. Schematic diagrams, showing topography and structure of the axial zone of ridges of varying spreading rate. (A) Slow-spreading ridge: pronounced axial valley, faulting, discontinuous volcanism, exposure of detachment fault, and underlying "megamullions." (B) Intermediate-spreading ridge. Subdued inner wall of axial valley, more continuous central volcano. (C) Fast-spreading ridge, little or no axial valley, smooth surface, and a narrow summit rift in a continuous central shield volcano. After MacDonald (1982). Redrawn after Twiss and Moores (2007, fig. 19.15, p. 594). Vertical exaggeration, ~2:1.

(see Fig. 4B for locations). These windows provide a view of a number of common features: Pillow lavas tend to dip toward the spreading axis up to 40°–50°. Dikes are evident below the pillows, with the contact being complex and gradational. Dikes tend to dip ~40°–50° away from the spreading center. These dips are thought to represent the product of the oceanic intrusive-extrusive pile, providing accommodation space for extrusion of successive lenses of lavas at the spreading center. Faulting and brecciation are common in these exposures. Some less fractured dikes crosscut the highly fractured exposures. Gabbros are present in some places. In the Pito Deep, in situ magnetic measurements indicate that the direction of intrusion of the gabbro was horizontal, parallel to the ridge direction (Varga et al., 2008).

Propagating rifts also are a feature of ridges, especially fast-spreading ones. Along the East Pacific Rise are many places where the ridge crests of segments propagate or die back with time. The process produces chevron-like discontinuities in magnetic anomalies called *pseudofaults* by some workers (Fig. 10).

The oceans are very large, and the few exposures are small and scattered. For most of the oceans, we simply do not know exactly what underlies the ocean bottom. In addition, the extant oceans are younger than most of Earth's history. Thus we are left with indirect means of inferring what the oceanic crust really does look like.

Ophiolite complexes (the word comes from two Greek words: *ophis,* meaning snake, and *lithos,* meaning rock) provide some insight. Ophiolites are models for ocean crust and mantle formed by sea-floor spreading. These features are mafic-ultramafic complexes found in deformed belts throughout the world. Where best preserved, they display a sequence from bottom to top of tectonite, peridotite, commonly harzburgite or lherzolite, and dunite with included concentrations of chromite, ultramafic-mafic cumulates, massive gabbro-diorite-plagiogranite, a

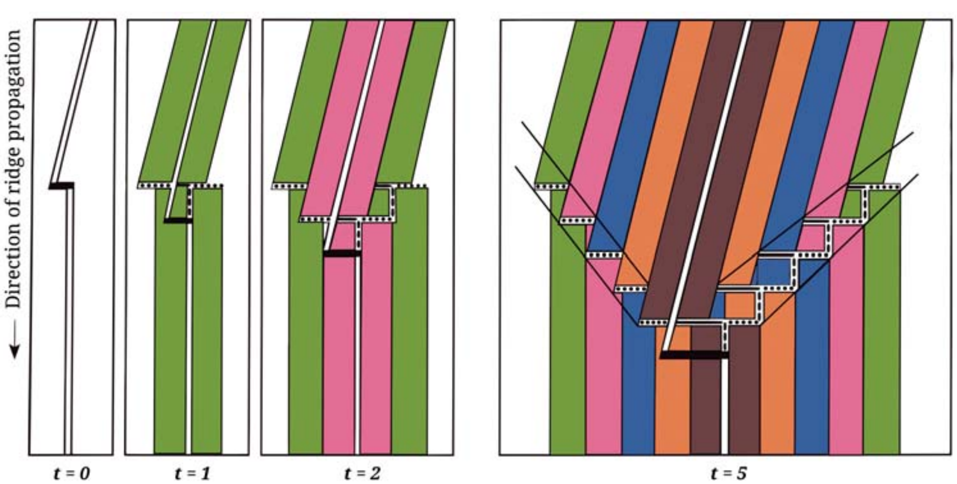


Figure 10. Schematic diagram, showing the effect of a propagating ridge on the pattern of magnetic anomalies for times t = 0, 1, 2, and 5. After each increment of spreading, the northern segment of the ridge has propagated southward, creating diagonal discontinuities called *pseudofaults.* After Hey (1977; see also Moores and Twiss, 1995, p. 102).

mafic sheeted dike complex with many half-dikes indicating a dike-within-dike formation, massive pillowed extrusive lavas, and an overlying sequence of pelagic or volcaniclastic sediments. (Fig. 11).

The importance of ophiolites lies in the fact that extant oceanic crust is hard to observe, and crust older than ca. 180 Ma has disappeared. Thus ophiolite complexes provide not only a window into modern ocean crust, but for ocean crust older than 180 Ma, *ophiolites provide the only information*.

Not all ophiolite complexes exhibit the complete sequence. Ophiolite with a complete sequence may represent a magma-rich, or fast-spreading ridge (Fig. 11A), Ophiolites formed at a magma-starved or slow-spreading ridge present incomplete or faulted sequences with serpentinite variably overlain directly by extrusive rocks and sediments (Fig. 11B). In intra-arc or back-arc environments, a complex overlap of deep to shallow intrusive and extrusive rocks may be present (Fig. 11C) . Hotspots or oceanic plateau may present a thick extrusive or intrusive section (Fig. 11D).

Ophiolites thus described are present throughout the world in deformed regions less than ca. 1000 Ma. Older deformed belts contain complexes thought to be equivalent to ophiolites, but a thicker magmatic section commonly has prevented a good exposure of the tectonized ultramafic layer or lithospheric mantle (e.g., Moores, 2002).

Many ophiolite complexes also contain half grabens of dikes and pillow lavas (Figs. 12, 13). One example, the Vourinos complex, in northern Greece, may represent a preserved spreading center with an amagmatic flank on one side (Fig. 12). The Troodos complex, Cyprus, exhibits evidence of magmatic healing of fault structures (e.g., Hurst et al., 1994; Granot, et al., 2011).

The oceanic crust probably is considerably more complex than the simple layered seismic model implies. Evidence from "tectonic windows" in the ocean floor (Karson, 1998) shows that even the ophiolite models displayed in Figures 11–13 are over-simplifications of the relations in observed exposures. Hawkins (2003) summarized the complexities of the marginal basins of the western Pacific and elsewhere. Paleomagnetic evidence shows that some magmas in oceanic crust have been intruded horizontally (Varga et al., 2008).

The exact relationships between classic ophiolite structures and those of mid-oceanic ridges are complex and not quite clear, in view of the observations from the oceans, mentioned above. It is not clear whether the directly exposed sections in oceanic regions or ophiolites are anomalous with respect to the main part of the oceanic crust. Also it may be that there has been too much emphasis on ideal ophiolite sequences and less on the more widespread exposures of partial sequences or faulted sequences. Perhaps the latter are more representative of the oceanic crust than heretofore realized.

Many petrological and geochemical studies have found that many ophiolite complexes have boninitic compositions (high Mg and Si and depleted trace elements; e.g., Miyashiro, 1975; Pearce, 1980; Bloomer et al., 1995; Shervais and Kimbrough, 1984; Bédard and Hebert, 1992), indicating formation from mantle depleted of its more primitive constituents. Such compositions are also found in modern backarc basins and forearc regions and has led many geochemical workers to argue that all ophiolites are derived from such regions (e.g., Pearce, 1980). Other workers have argued that mantle heterogeneities and historical contingency allow such ophiolites to have formed in mid-oceanic ridge settings (Moores et al., 2000). The stratigraphy on top of an ophiolite has also been emphasized by some as indicating a mid-oceanic setting (Hopson et al., 2008). Recent discovery of high pressure minerals in the Luobusa ophiolite, Tibet (Yang et al., 2007), indicates derivation of some chromitite by solid flow of mantle rock from depths as great as 380 km. This fact, and the chromite's age as greater than that of the ophiolite itself, indicate a deep mantle involvement, regardless of the lava compositions and tectonic locations.

Nevertheless, ophiolite complexes, as strictly defined, mostly represent tectonic fragments of oceanic crust formed at some sort of spreading center, either in a mid-ocean or supra-subduction-zone setting. These complexes thus provide insight into possible tectonic and magmatic processes that produce oceanic crust. And, as mentioned above, they provide the only insight into oceanic spreading processes for times prior to the oldest oceanic crust.

TRANSFORM FAULTS, STRIKE SLIP FAULTS, AND RELATED FRACTURE ZONES

Transform faults are one of the three principal plate boundaries, whereby two lithospheric plates move past each other with no creation or destruction of lithosphere (Fig. 14). Most transform faults are between offset ridge segments. These ridge-ridge transform faults are those first recognized by Wilson (1965).

Transform faults are possible between each of the three principal plate boundaries. Ridge-ridge transform faults tend to remain constant in length; earthquakes are limited to the transform offset between the ridge segments. The transform faults continue, however, as long scars or fracture zones beyond the ends of the active parts of the faults.

A ridge-trench transform fault separates a ridge from a subduction zone. On the present Earth, transform faults of this type include those bordering the Caribbean plate: the Cayman fracture zone east and west of the Cayman trough, the Bocono–El Pilar fault and its continuations along the northern coast of South America, and the faults separating the South Sandwich plate from the South Atlantic and Antarctic plates. Faults of this type vary in length, depending upon whether the plate adjacent to the spreading center is the overriding or downgoing plate. In the former, the transform fault lengthens at half the spreading rate. In the latter case, the transform faults change length by the half rate of spreading minus the subduction rate.

The Dead Sea transform fault (at the east end of the Mediterranean) is a left-lateral fault that separates the African plate from the Arabian plate and transforms the motion of the spreading

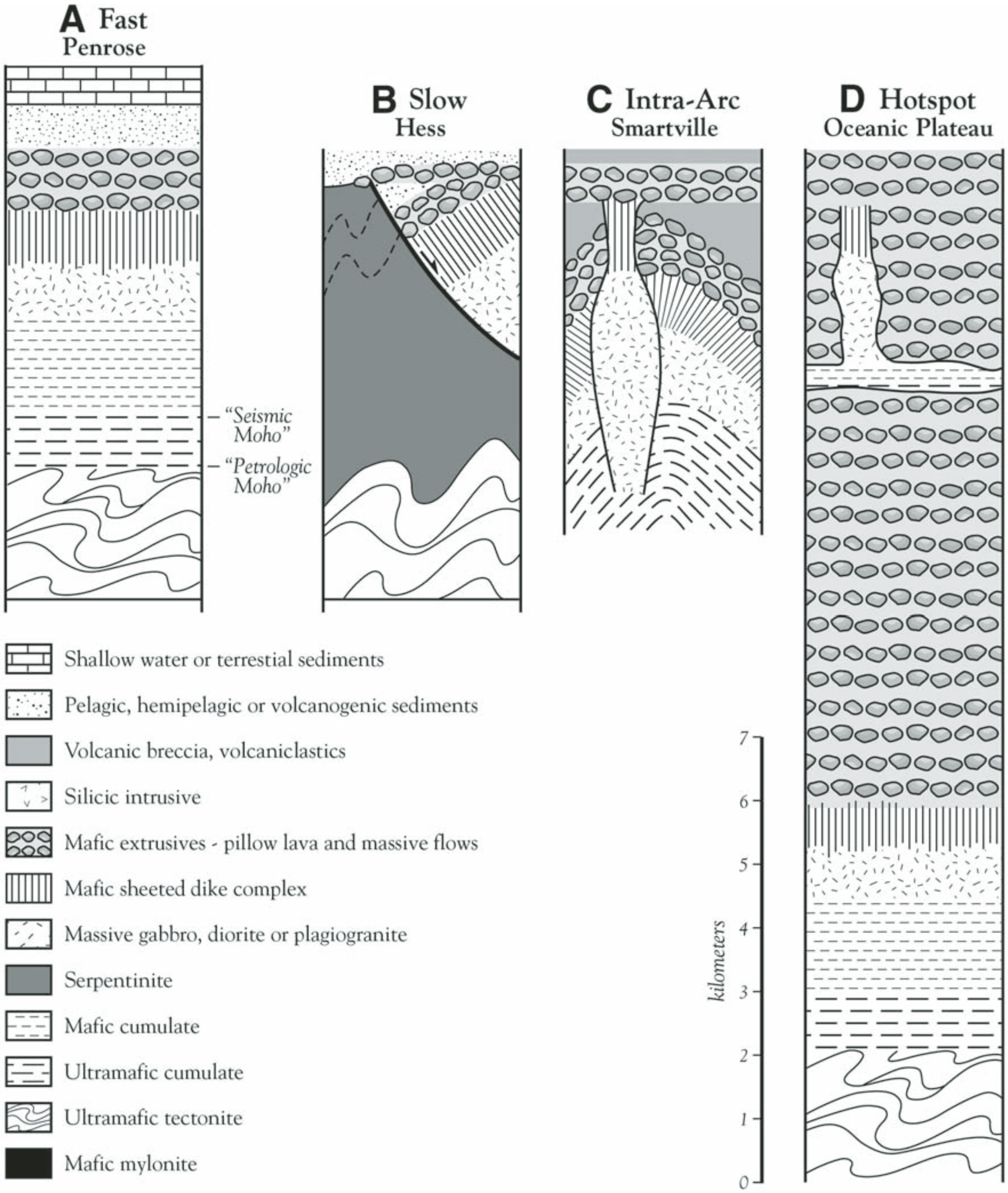


Figure 11. Schematic columnar sections of ophiolite types and inferred equivalent oceanic crust type. (A) Complete ophiolite sequence, according to the Penrose Conference definition (Anonymous, 1972), symbolic of a magma-rich (generally fast-spreading) center. (B) Faulted, incomplete sequence, possibly characteristic of a magma-starved spreading, a so-called *Hess-type* ophiolite (cf. Fig. 1 and Fig. 8). (C) Complex composite section of oceanic-island-arc spreading-center sequences, designated as *Smartville type* for the Smartville Complex in the northwest Sierra Nevada, California (e.g., Dilek et al., 1991). (D) Possible hotspot (oceanic plateau) section of oceanic crust. Note small high-level pluton (after Moores, 2002, fig. 1).

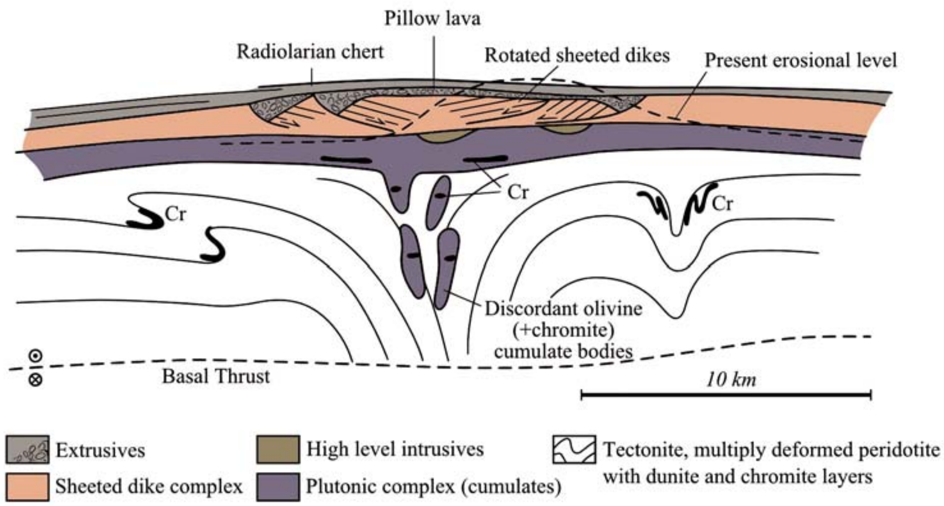


Figure 12. Schematic cross section of Vourinos complex, northern Greece, interpreted as a fragment of oceanic crust and mantle preserving a faulted and eroded fossil spreading center. After Moores (2003, fig. 2B, p. 25).

Red Sea into the intracontinental Bitlis-Zagros subduction zone, extending from Turkey into Iran.

Transform faults also can separate two trenches. When separating two subduction zones of the same polarity and on an overriding plate, the transform fault length remains constant. When separating subduction zones of opposing polarity and dipping toward each other, the transform fault lengthens at a rate equal to the subduction zone rate. Faults of this type include the Alpine fault, New Zealand, as is the complex fracture zone between the Vanuatu and Tonga-Kermadec subduction zones, as well as the intracontinental Chaman fault, Pakistan-Afghanistan, separating the Chaman subduction zone south of Pakistan from the Karakorum-Himalaya intracontinental subduction zone. If the subduction zones dip away from each other, the transform fault shortens at the rate of subduction.

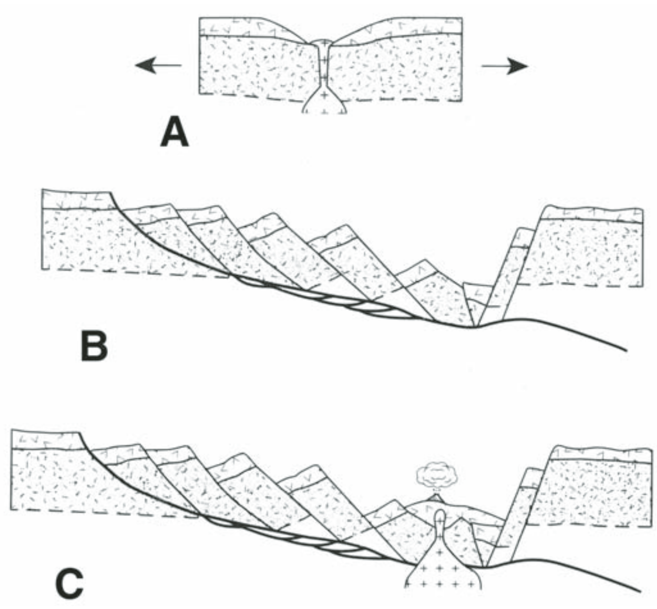


Figure 13. Schematic cross section of dike-extrusive complex of part of the Troodos ophiolite complex, Cyprus, showing tilted fault blocks deformed on a listric normal fault, bottoming out in a subhorizontal detachment fault representing the brittle-ductile transition in oceanic rocks. (A) Extension by magmatic accretion. (B) Extension by structural thinning. (C) Extension by renewed magmatic accretion. After Varga and Moores (1990).

Transform faults also can separate two triple junctions. The San Andreas–Gulf of California (Sea of Cortez) system, western North America, separates two triple junction zones that are moving away from each other, the Mendocino and the Rivera triple junctions (e.g., Atwater and Stock, 1998; McKenzie and Morgan, 1969). In this case, the fault length changes in a complex way depending on the nature of the triple junctions. In the San Andreas–Gulf of California situation, the fault lengthens by the sum of its displacement rate plus the half-spreading rate of the East Pacific Rise.

Whether continental or oceanic, transform fault geology is complex. In the oceans we have abundant evidence only from ridge-ridge transform fault zones. The width of the fault zone separating ridge segments ranges from ~10 km for slow spreading ridges up to >100 km for fast-spreading ridge offsets (Fig. 15). Sharp ridges and troughs particularly characterize slow-moving transform faults. Recovered rocks include serpentinites; metamorphosed, mylonitized, and foliated ultramafic rocks; mafic plutonic rocks; and metavolcanic rocks. Some coarse talus breccias interfinger with pelagic sediments. Serpentinite diapirs are possibly present along some transform faults (e.g., MacDonald et al., 1979). In fast-slipping oceanic transform faults, small subordinate spreading centers may be present. The ridge-transform intersection is a complex zone of simultaneous igneous intrusion-extrusion and deformation.

The schematic maps in Figure 15 show slight "j-shaped" or "reverse j-shaped" curves in the magnetic anomalies. These deflections of the anomalies and the accompanying topography (abyssal hllls) appear to be the result of deformation of the transform zone and the fracture zones beyond the active transform

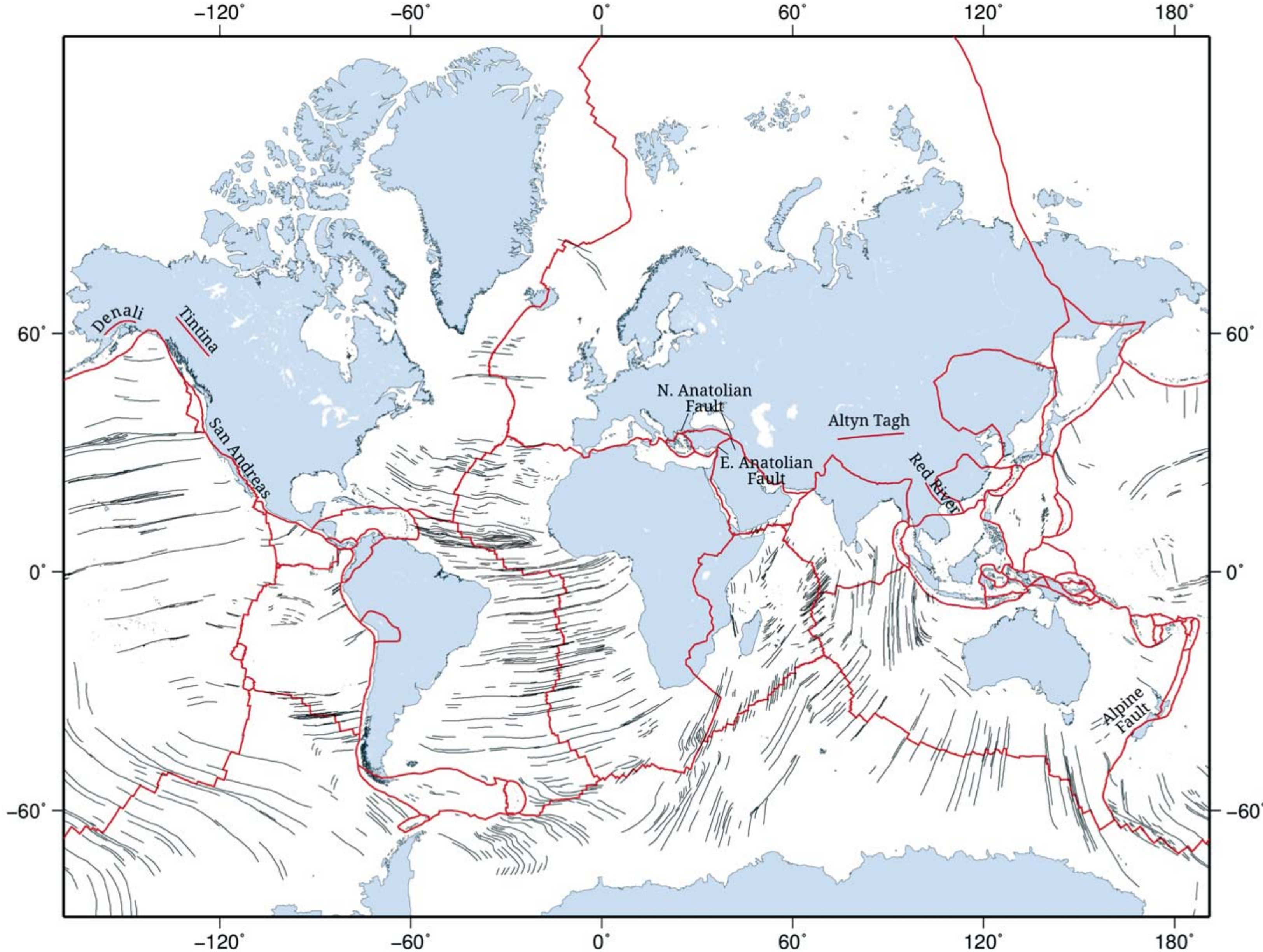


Figure 14. Mercator projection of the Earth's surface, showing plate boundaries (red lines), fracture zones (black lines), and large continental strike-slip faults (labeled red lines) of the world. The curves at the ends of some fracture zones reflect changes in the location of relative poles of spreading during evolution of the ridge-transform system. After Bird (2003) and Matthews et al. (2011).

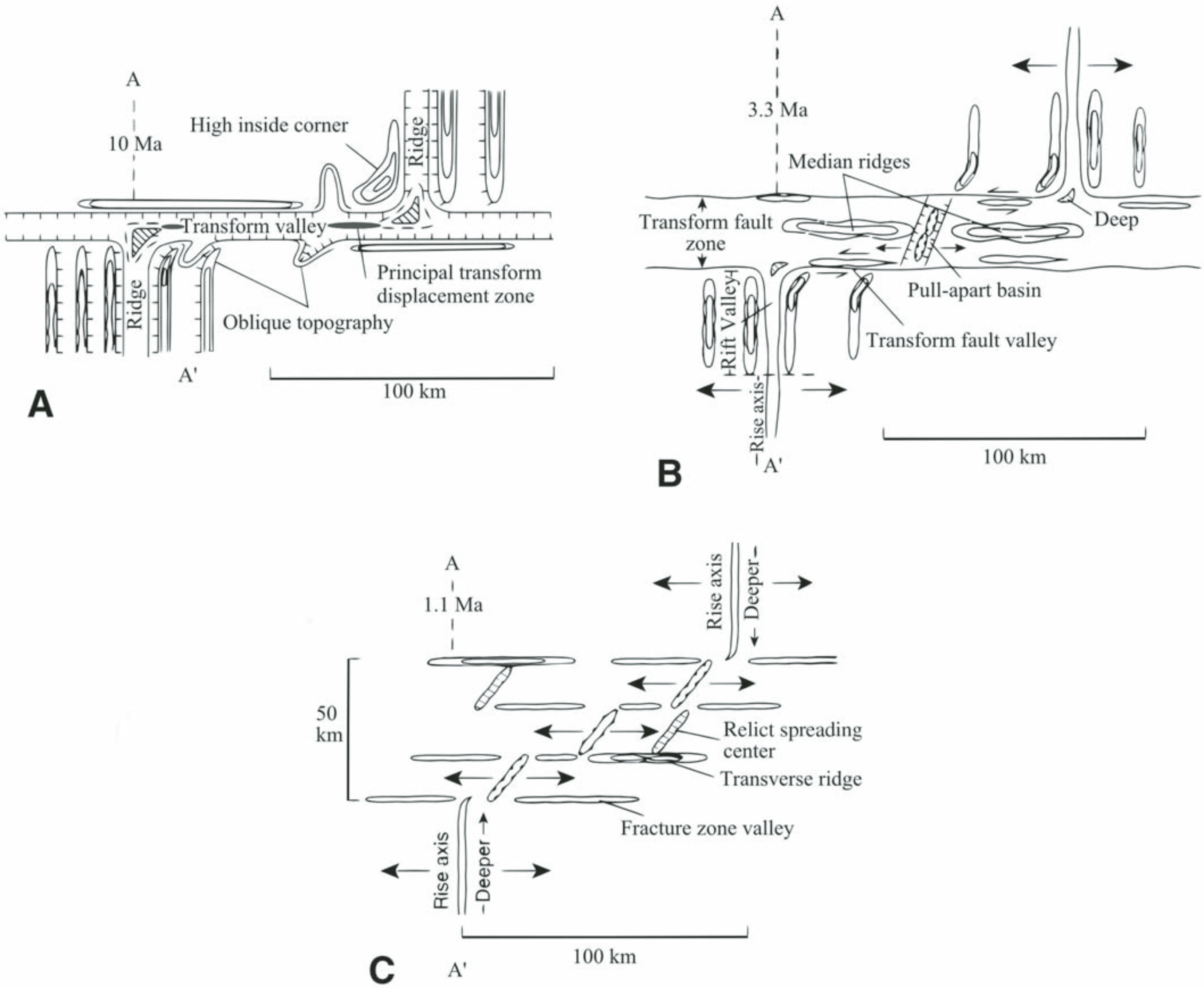


Figure 15. Schematic topographic maps, showing varying physiography of oceanic transform faults, depending on rate of slippage. (A) Model for slow-spreading center (full spreading rate <5 cm/a). Note rifted ridge, high inside corners, relatively narrow transform valley, and a principal transform displacement zone. (B) Intermediate spreading center physiography (full spreading rate 5–9 cm/a). Wider transform fault zone, including parallel ridges and valleys, one or more pull-apart basins, and less pronounced ridge-flanking topography. (C) Fast-spreading physiography, showing a complex transform fault zone tens of kilometers wide including transverse ridge and valleys, several active or inactive pull-apart basins, and distributed shearing. Note "j" shaped ridges at spreading center–ridge axis intersections. After Fox and Gallo (1984, figs. 2 and 4). See also Twiss and Moores (2007, fig. 19.23, p. 604).

fault. Such relative movement may result from change in pole of rotation of the plates and resultant deformation along the transform fault and fracture zone either by rotation, relative movement and compression (so-called transpression), or extension (so-called transtension) (Croon et al., 2010).

Active on-land transform faults are also wide zones. The San Andreas transform fault system (Fig. 16B) consists of a complex set of subparallel faults, folds, thrust faults, and pull-apart basins, extending over 1000 km from the obliquely rifting Gulf of California, past the restraining bend in southern California with neighboring rapidly uplifting mountains, to the long NW-trending central to northern Coast Ranges with a family of faults, and also to scattered active and inactive pull-apart basins, with associated volcanism, alternating with folding.

The Alpine fault of New Zealand (Fig. 16A; Carter and Norris, 1976; Wells et al., 1999) is also a wide multiple fault zone, extending more than 500 km between two subduction zones and consisting of a large family of related faults. The fault has a long, complex history related to interaction between the Antarctic, Australian-Indian, and Pacific plates.

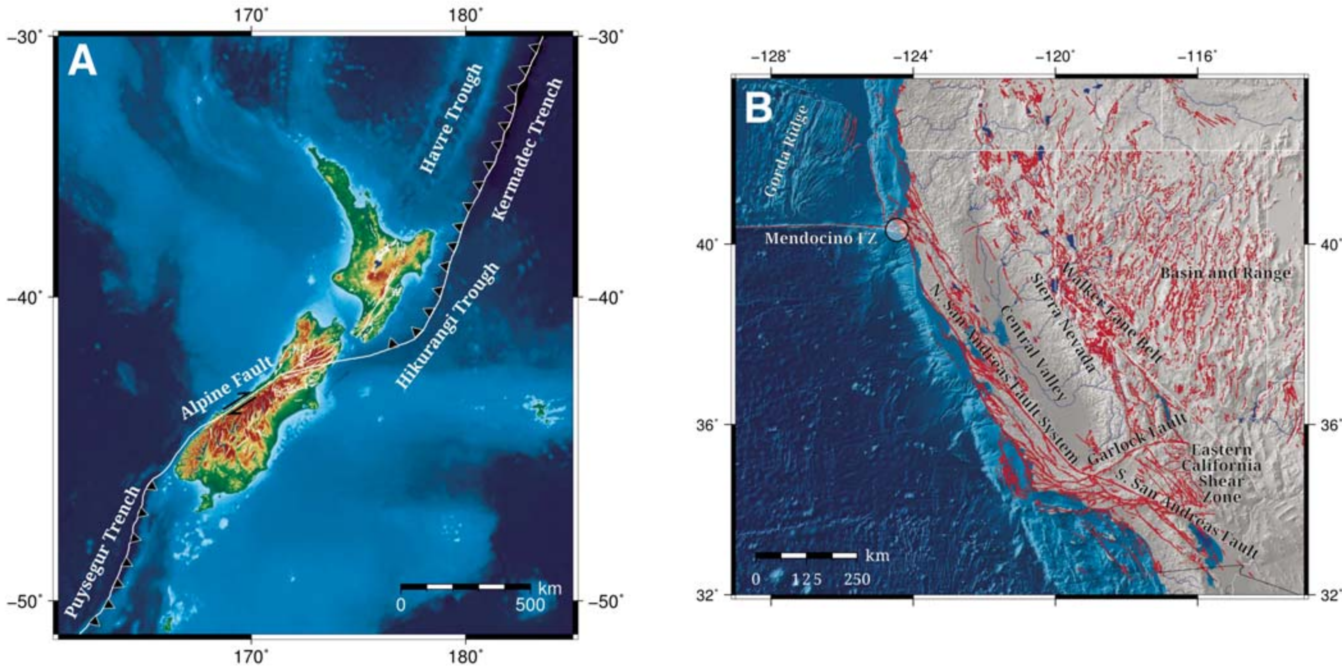


Figure 16. On-land transform faults. (A) Alpine fault, New Zealand, a trench-trench transform fault. Redrawn after Bird (2003); see also Moores and Twiss (1995, fig. 6.25, p. 150). White lines are subsidiary faults, after New Zealand Geological Survey. (B). The San Andreas fault system, and related fault systems. Quaternary faults in red. Relief after Amante and Eakins (2009). Faults after USGS Quaternary Fault Map (U.S. Geological Survey and California Geological Survey, 2010).

The strength of transform faults has been a subject of recent work. For example, Platt et al. (2008) argued on the basis of modeling and experiments that a transform fault is a relatively weak shear zone that exerts a small amount of stress on the bordering rock.

The strength of the San Andreas fault has long been controversial. A weak fault model has been proposed, based on low heat flow (Lachenbruch and Sass, 1980, 1992), inferred orientations of principal stress directions (e.g., Hickman and Zoback, 2004; Zoback, 2000; Zoback et al., 1987), and numerical models (e.g., Carena and Moder, 2009; Bird and Kong, 1994). Other workers (e.g., Scholz, 2000; Melosh, 1996; Andrews and Ben-Zion,1997) argued for a strong fault, with low heat flow attributed to groundwater circulation or dynamic weakening during rupture. Core samples obtained by the San Andreas Fault Observatory at Depth (SAFOD) from the fault's central creeping section showed the presence of the very weak clay mineral saponite in the fault itself (Lockner et al., 2011), but also a sharp increase in strength in the wall rocks a short distance from the fault (Carpenter et al., 2012). The San Andreas clearly is a very weak fault at this location, but its strength at other localities, especially along its northern and southern locked sections, is still questionable.

There is no real evidence for the strength of oceanic transform faults. From studies of the Troodos ophiolite, Cyprus, Abelson et al. (2002) suggest that that transform fault was unexpectedly strong relative to the strength of the neighboring spreading center faults.

Recognized on-land examples of intraoceanic transform faults also mainly comprise those separating spreading centers. The South Troodos transform fault, Cyprus (Moores and Vine, 1971; Gass et al., 1994), displays a complex zone of faults, interlayered talus breccias and pelagic or metal-rich sediment, and deformed meta-igneous rocks (Fig. 17A), passing downward into foliated harzburgite tectonite. The Coastal Complex next to the Bay of Islands complex, Newfoundland, displays a similar complex geology, but at a deeper level (Fig. 17B). The complex rocks are exposed at a level originally a few kilometers deep (Karson and Dewey, 1978; Karson, 1984; Nicolas, 1989). The "type" ophiolites of Italy and the western Alps (Fig. 17C) may include a transform fault type of oceanic crust. Exposures consist chiefly of serpentinized peridotite overlain by sedimentary breccias and/or pillow lavas, and pelagic sediments. Mafic intrusive and dike complexes are mostly absent. Some exposures of the Apennine ophiolites, however, may also represent slow-spreading oceanic crust. Abbate et al. (1981) portray the ophiolites as representing the remnants of a complex spreading center–transform system, analogous to the Gulf of California. (The ophiolite complexes of the Dinarides and Hellenides, shown in Figure 17C, mostly formed by spreading processes.) The New Caledonia ophiolite (Fig. 17D) is principally composed of peridotite and its metamorphic sole. Peridotite fabrics in two regions, (1) a 15-km-thick shear zone on the Bogota peninsula, and (2) the Belep shear zone on NW New Caledonia and neighboring islands, may represent remnants of transform faults, separating spreading centers in the ocean basin to the north. Subsequent ophiolite emplacement on the Norfolk Ridge, a narrow continental ridge east of Australia, possibly was controlled by these preexisting fractures (Titus et al., 2011; Nicolas, 1989).

These four on-land transform fault exposures show transform fault characteristics and activity at various depths. The South Troodos transform (Arakapas) fault and the Italian ophiolites display surface and near-surface features. The Coastal Complex, Newfoundland, shows transform faulting in lower oceanic crustal levels. Both the South Troodos and the New Caledonia ophiolite show evidence of transform faulting within the mantle.

No information is available as to the geology of transform faults separating oceanic subduction zones. A fault zone separating the northern Sierra–southern Klamath Mountains of northern California may represent an on-land, trench-trench transform fault, separating two subduction zones dipping in the same direction, toward each other, or away from each other (e.g., Dilek et al., 1991; Moores et al., 2002). The fault may have been active in Middle Jurassic time (ca. 160 Ma).

Major intracontinental strike-slip or transform faults include the Karakorum and Altyn Tagh faults in central Asia; the North Anatolian and East Anatolian faults in Turkey; the Red River fault, southeast Asia; the Chaman Fault, Pakistan-Afghanistan; and fault zones separating the Alboran plate from the Iberian Peninsula in Morocco (see Fig. 14) . The Great Glen fault of the northwestern UK is thought to be a large-displacement strike-slip (or transform) fault of late Paleozoic age. The Queen Charlotte Islands fault is an active oceanic transform fault separating the North American and Pacific plates west of Canada.

Active strike-slip faults in central-eastern Asia constitute part of the complex collision zone between the Arabian, Indian, and Eurasian plates (see Fig. 4). For example, the central Altyn Tagh fault is moving at ~8–12 mm/a (Gold et al., 2011).

Molnar and Dayem (2010) examined some 40 active intracontinental faults in the Americas, Asia, and southeast Asia with rates of motion of 10 mm/a or more. They concluded that many of these faults lie adjacent to relatively strong regions—mature oceanic lithosphere or Precambrian shields. Thus the faults may represent a concentration of intraplate motion against these strong areas.

CONVERGENT MARGINS (SUBDUCTION ZONES)

Convergent margins, or subduction zones, constitute plate margins where plates move toward each other, with the motion accommodated by one going down beneath the other (Fig. 4). Convergent margins comprise some 45,000 km of plate boundary

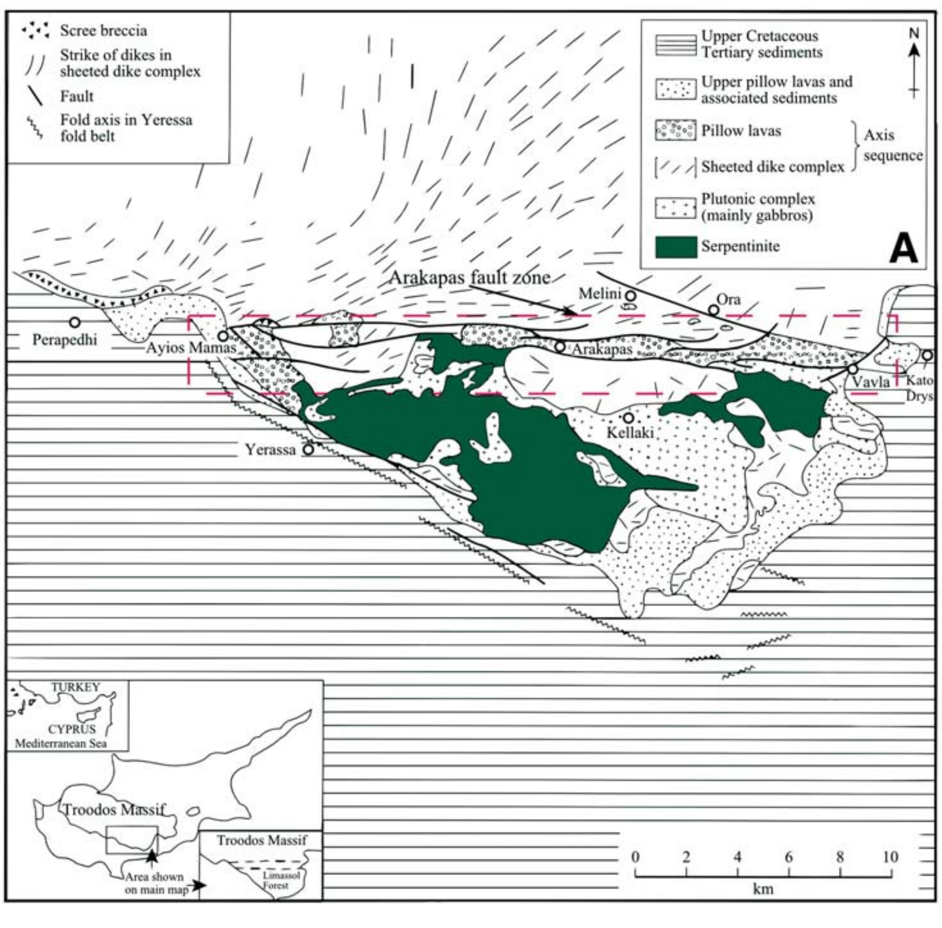


Figure 17 (*continued on following page*). Possible oceanic transform faults exposed on land. (A) Arakapas transform fault, Troodos complex, Cyprus (Redrawn after Moores and Vine, 1971; Simonian and Gass, 1978).

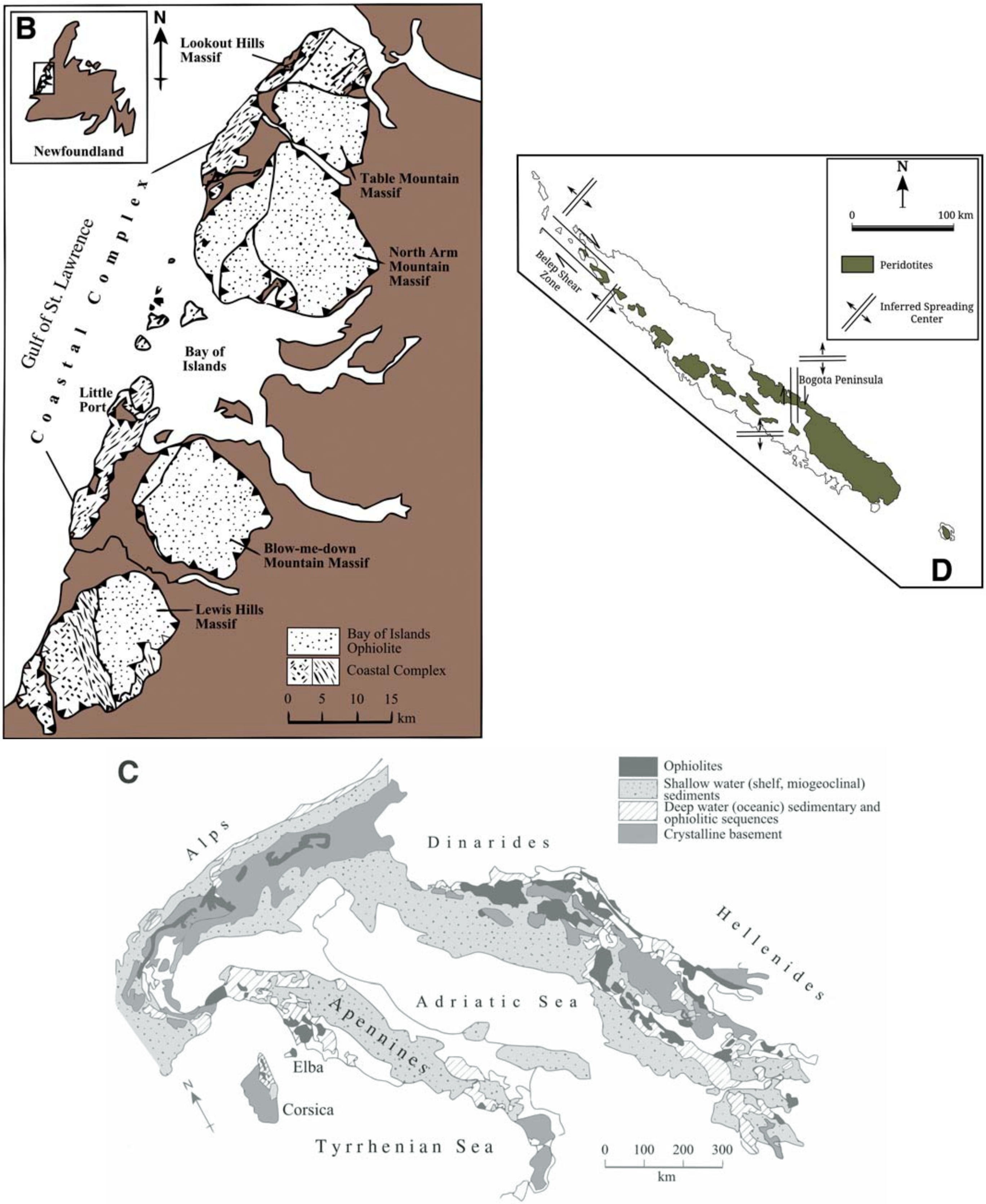


Figure 17 (*continued*). (B) Coastal Complex, Bay of Islands, Newfoundland, redrawn after Karson (1986). (C) Italy-Alps, redrawn after Abbate et al. (1980). See also Moores and Twiss (1995, p. 138–141). (D) New Caledonia, showing the Bogota Peninsula and the Belep shear zones, two possible oceanic transform fault zones (after Titus et al., 2011; Nicolas, 1989; Prinzhofer and Nicolas, 1980).

on the Earth. They consist of two principal types, *oceanic* (or island) arcs (e.g., Larter and Leat, 2003), where two oceanic plates converge, and one descends beneath the other, and *continental* arcs, where an oceanic plate descends beneath a continental plate. A complication is that in some regions, specifically Taiwan and New Guinea–Indonesia, a continent (Asia and Australia, respectively) are on the downgoing plate, and in some regions, specifically in the Alpine-Himalayan belt, a continent is descending beneath another continent along a continental subduction zone. These examples are of collisions, discussed in the following section.

Both oceanic and continental arcs contain similar general features (see Fig. 18).

A. An outer ridge or swell, where the downgoing plate bends into an inclined position to begin its descent into the mantle. This swell represents an elastic bending of the downgoing plate. Minor normal faulting, earthquakes, and even volcanism characterize these flexures (Hirano et al., 2006).

B. A boundary between the two plates. This boundary is where the plates meet, and the descending plate plunges into the mantle. The interface is the subduction zone. A topographic trench marks this boundary where it crops out on the Earth's surface, with negative free air gravity anomalies. The existence of negative gravity anomalies associated with these trenches was first known by work of Vening Meinesz and associates in the 1920s and 1930s (e.g., Vening Meinesz, 1955; Hess, 1939), but their true significance did not become evident until the Plate Tectonic Revolution of the late 1960s and early 1970s. Trenches are the deepest features in the ocean. The average slope of both sides of a trench is ~10°–20° or so. Depending upon the rate of subduction and the supply of sediment to the forearc regions, there may be a layer or turbidite in the axis of the trench.

C. **An accretionary prism.** On many subduction zones, there is a *pileup* of material along this boundary derived from the downgoing plate and added to the overriding plate. This accretionary prism comprises a series of imbricate thrust faults and intervening basins wherein sediment is received from both sides and is tectonically intermixed.

The bottom thrust of an *accretionary prism* is a *décollement,* i.e., a detachment horizon between the accreting material and the downgoing plate. Fluids, principally water, lubricate this zone, diminishing the friction between the two plates and aiding in the subduction process.

Accretionary complexes dominantly consist of a series of thrust faults dipping in the same direction as the subduction zone itself. These thrusts generally are progressively younger toward the trench. The accretionary complex develops a *critical taper,* which is to maintain an equilibrium shape of the accretionary prism that is a function of the slope of the downgoing plate, the upper surface of the accretionary material, friction along the basal décollement, and the strength of the rocks themselves (Davis et al., 1983; Dahlen and Suppe, 1988).

Mélanges (chaotic mixtures of rocks of diverse origins) also are characteristic features of accretionary prisms. These features

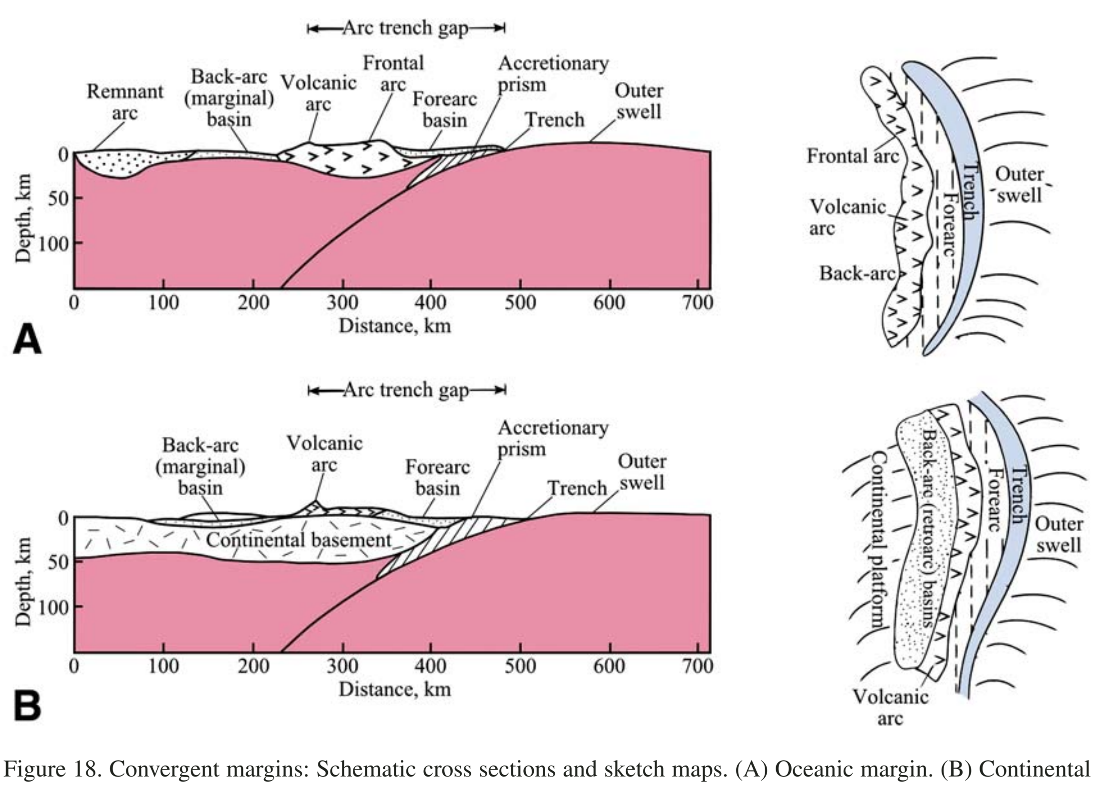


Figure 18. Convergent margins: Schematic cross sections and sketch maps. (A) Oceanic margin. (B) Continental margin. Redrawn after Moores and Twiss (1995, fig. 7.2A, B).

may be of sedimentary or tectonic origin, or both (e.g., Hsü, 1968; Wakabayashi, 2012).

Recently, several active accretionary prisms have been instrumented by the IODP (Saffer and Tobin, 2011). Data from these instrumented accretionary complexes indicate an interplay between fluid pressure, fluid production by dewatering, compaction, metamorphic reactions, tectonic overpressure and rock strength affecting the earthquake behavior of the subduction zone. Some zones of elevated fluid pressure may be imaged in seismic reflection profiles (Saffer and Tobin, 2011).

Many subduction zones apparently do not possess accretionary prisms. Sediments on the downgoing plate apparently proceed into the mantle without accumulating along the plate edge. Some subduction zones even appear to be eroding (Von Huene and Scholl, 1991; Clift and Vannucchi, 2004). Tectonic erosion apparently occurs in subduction zones with high rates of convergence and/or thin sedimentary cover. In these cases, sediments go into the mantle or they are underplated on the bottom of the overriding plate (Clift and Vannucchi, 2004).

Seismic activity associated with subduction zones goes as deep as 700 km. Earthquakes at higher regions are near the top of the subducting slab, whereas those deeper are mostly in the slab itself (e.g., Isacks and Molnar, 1972). As schematically depicted in Figure 2, seismic tomography shows that slabs can penetrate into the lower mantle.

Subduction zones produce the largest earthquakes on Earth (Moore, 2007). Drilled subduction thrusts exhibit 20–40-m-thick zones of scaly clay. Fracture porosity and permeability vary, and dilation by high fluid pressure may have occurred. Progressive underplating of underthrust sediments also is evident. Seismic events begin at ~125 °C (Moore, 2007). On-land examples of seismogenic faults include zones of rock formed by frictional melt—fluidized but unmelted rocks—and incipiently metamorphosed fault rocks (Moore, 2007). In some obliquely subducting zones the motion is evidently partitioned into dip-slip and strike-slip components. Stress determinations in a hole in the Kumano forearc basin of the Nankai subduction zone, Japan, display a maximum horizontal stress approximately parallel to the subduction direction, and the stress field apparently partitioned into dip slip and strike slip components (Lin et al., 2010).

Metamorphic rocks indicating formation at high pressure, low temperature conditions are abundant in accretionary complexes (e.g., Ernst, 1970, 1975; Platt, 1986; Wakabayashi and Unruh, 1995). These rocks must have formed at high pressures in subduction zones; their exposure at the surface has been ascribed to some sort of counterflow (Ernst, 1970, 1975) or detachment faulting within the accretionary prism (Platt, 1986).

D. An "arc-trench" gap lies between the trench–accretionary prism and the active volcanic arc (Dickinson, 1970b). At some locales this gap is the locus of a sedimentary basin (e.g., Ingersoll, 2012), whereas at others it is a sediment-free plateau of older material (von Huene and Scholl, 1991).

E. An active volcanic arc lies over a region where the earthquakes are ~125–150 km deep (Dickinson and Hatherton, 1967). Thus the width of the arc-trench gap varies with the dip of the subduction zone. In some places where the plate dips at a shallow angle, the gap is wide (e.g., Peru-Chile, southern Alaska). In others, the plate dips steeply, and the gap is narrow (e.g., Marianas). Active volcanoes, chiefly andesitic, characterize this arc, with the volcanoes typically spaced ~80 km apart (Miyashiro, 1975; Marsh and Carmichael, 1974; Marsh, 1979).

Arcuate chains of active volcanoes characterize both oceanic and continental arcs. Examples of oceanic arcs in the Pacific are the Kurile Islands, the Japan Islands, the Ryukus, the Izu-Bonin-Marianas; the Bismarck-Solomons-Vanuatu, Indonesia, Tonga-Kermadec; and the Aleutians; in the Atlantic the Greater-Lesser Antilles and the south Scotia arc; in the Mediterranean the Calabrian and Hellenic arcs; and in the Indian Ocean the Indonesian arc.

Most arc axes have volcanoes rising above the frontal arc or continental platform. Oceanic arcs tend to have a uniform elevation of volcanoes above sea level, whereas continental arcs have volcano heights a uniform distance above the continental platform. For most continental arcs, the platform is near sea level. The platform of the Andes, however, lies some 3–6 km high. The reason for this high platform is unclear (e.g., Hamilton, 1969b; Hildebrand, 2013).

Intra-oceanic arcs display a crustal thickness of ~25–30 km, with normal oceanic crustal thicknesses on either side. The basement of most island arcs in the Pacific is Eocene or younger oceanic rocks, but in Japan the rocks are as old as Paleozoic, and the Japan arc apparently has been active since Paleozoic time. Japan represents an arc that separated from Asia in Neogene time and may now be undergoing reversal (e.g., Ogawa et al., 1989).

Examples of active continental arcs include those around the Pacific: Kamchatka, the Alaskan peninsula, southern Alaska, the Cascades, the Trans-Mexico volcanic belt, Central America, and the Andes. Most of these arcs display crustal thicknesses of 25–30 km, with arc platforms of ~0–1000 m elevation. The Andes, however, are an exception, as crustal thicknesses there are as great as 50–70 km, and, as mentioned above, platform elevations of 3–6 km. Major regions of batholithic exposures of granitoid rocks characterize the East Pacific margin, and parts of eastern Asia. These batholiths are also thought by most workers to be the result of subduction-related plutonic activity (e.g., Dickinson, 2004, and many others).

F. The "backarc region" also varies. Behind most oceanic arcs, it is a zone of extension, and there is an active basin where slow sea-floor spreading is taking place. Most circum-Pacific continental arcs display a backarc region of little or no tectonic activity. In the Andes, however, there is an active fold and thrust belt in the backarc region. This region has been used as a global example for continental arcs (e.g., Kay and Ramos, 2006).

Backarc basins have a number of origins (e.g., Uyeda and Kanamori, 1979). They can be entrapped oceanic basins, such as the Aleutian and west Philippine basins; basins formed by spreading, such as the Lau basin behind the Tonga Kermadec island arc; the Mariana trough behind (west) of the Mariana island arc; part of the Scotia Sea, behind (west) of the Scotia arc; and the Japan Sea, behind (west) of the Japan island arc. Basins also form along oblique "leaky" transform faults, such as the Gulf of California and the Andaman Sea. Extensional faulting apparently accompanies earthquakes in these regions (Uyeda and Kanamori, 1979). Other backarc regions, such as the Peru-Chile subduction zone, exhibit compression in the backarc region. The differences may have resulted from high compressive stresses perpendicular to the downgoing slab or to recent collisional histories, as outlined below.

Backarc basins may also be the product of slab rollback (e.g., Schellart et al., 2006), wherein as subduction proceeds, the slab not only falls into the mantle parallel to the seismic zone, but the slab bend itself moves away from the subduction zone as motion proceeds. This migration away from the subduction zone would cause it to move oceanward, and the island arc would keep up by migrating away from its backarc region. As mantle convection in the overriding plate probably follows the downgoing slab, the result would be to open up a backarc basin rearward of the arc.

In many cases in the western Pacific, opening of a backarc basin has taken place within the arc itself, leaving behind so-called *remnant arcs* (Karig, 1972). The southwestern Pacific, with its many complex island arcs, ocean basins, and a history of collisions (see below) may have had a history of back-basin opening, collision and arc reversal, and ophiolite obduction occasioned by slab rollback. (Schellart et al., 2006; Hall, 2002).

COLLISIONS

Collisions are zones where a continent, island arc, oceanic plateau, or other thick piece of crust on a downgoing plate arrives at a subduction zone. If the arriving crustal block is too thick, probably ~15 km for continental or island-arc crust, or 30 km for an oceanic plateau (Cloos, 1993), the buoyancy of the plate becomes less than that of the asthenosphere. The plate will cease its activity unless the denser, lower part is detached from its upper part. Eventually subduction ceases at this location, and the plate tectonic pattern undergoes a rearrangement of plates, of plate motion, or of plate boundaries. New plate boundaries develop, or increased movement occurs on existing plate boundaries. The western Pacific displays a rich history of collisions and "flip" in the subduction direction, accompanied by slab rollback (Hall, 2002; Schellart et al., 2006).

Collisions thus defined are an integral part of the orogenic process. However, because they are varied and not always recognized, it is appropriate to discuss them separately. In addition, there is widespread confusion about terminology. Both the literature within the Earth science community and the general public unfortunately refer to plate convergence as *plate collision.* Convergence or subduction, and collision, are not the same process. Convergence-subduction allows for smooth operation of plate tectonics, whereas a collision disrupts it and causes a rearrangement of plate motions or margins, or both. It is important to keep these two processes separate.

Seven possibilities for collisions are illustrated schematically in Figure 19. They include (A) active continent–passive continent, (B) active island arc–passive continent, (C) active continent–backarc, (D) oceanic forearc–backarc, (E) two active continents with subduction zones dipping away from each other, (F) oceanic forearc–active continent, (G) oceanic forearc–oceanic forearc, and (H) active arc–oceanic plateau. Only a few of these collisions are active at the present time.

Presently active collisions include the following (see Moores and Twiss, 1995, p. 212–245; Twiss and Moores, 2007, p. 626–639). (1) The Izu-Bonin island arc is actively colliding end-on with Japan along the Boso triple junctions (type D, Fig. 19; e.g., Ogawa et al., 1989); (2) in NE Japan the Kurile island arc is colliding with Japan in central Hokkaido (type D, Fig. 19; e.g., Tsumura et al., 1999); (3) the Asian plate is colliding with the east-dipping West Luzon subduction zone in Taiwan (type B, Fig. 19; e.g., Byrne and Liu, 2002); (4) the Sangihe and Halmahera island arcs are colliding with each other with opposing dip of subduction zones (type E, Fig. 19; McCaffrey et al., 1980; Cardwell et al., 1980; MacPherson et al., 2003); (5) the India-Asia collision (type A, Fig. 19) is a collision of an active continental margin (Tibet; south Asia), and a passive continental margin (the northern margin of India), pushing up the Himalaya; e.g., Yin and Harrison, 1996); (6) more generally, the Arabia-Eurasia collision is an active continental margin–passive continental margin collision (type A, Fig. 19), in which the Taurus-Zagros Mountains of Iran and Turkey reflect the active continental margin (Turkey, Iran, Afghanistan) colliding with the passive margin (Arabia), and the Greater Caucasus represents an early stage of this collision (Forte et al., 2012); (7) the Eratosthenes seamount–Cyprus collision is an active continental arc (Anatolia-Cyprus) and a passive margin (Eratosthenes seamount–Africa); (8) an ongoing collision exists between the Australian continent and the north-dipping Banda–New Britain subduction zone, with the polarity of subduction undergoing a change from north to south dipping in New Guinea (type A, Fig. 19; Schellart et al., 2006; Little et al., 2011); and (9) in addition, Hall (2002) and Schellart et al. (2006) document several recent arc-arc and arc-microcontinent collisions in the complex marginal basin region of the SW Pacific.

Each collision involves development of a suture, a scar of a disappeared ocean. The collisions also produce extensive fold-thrust belts that are mostly *synthetic,* i.e., with movement in the same direction as the subduction zone. When two opposing subduction zones collide, the relations become complex and difficult to decipher. It is even more difficult in ancient, now inactive, collisions. In the best cases the fold-thrust belts transport continental margin and shelf rocks toward continental interiors. Similar thrust features must accompany collisions of island arcs (McCaffrey et al., 1980), but the rock units are more difficult to separate from each other.

The fold-thrust belts of collision zones are in fact end points of many subduction accretionary complexes. Active accretionary complexes and active collision zones are the *actualistic,* modern

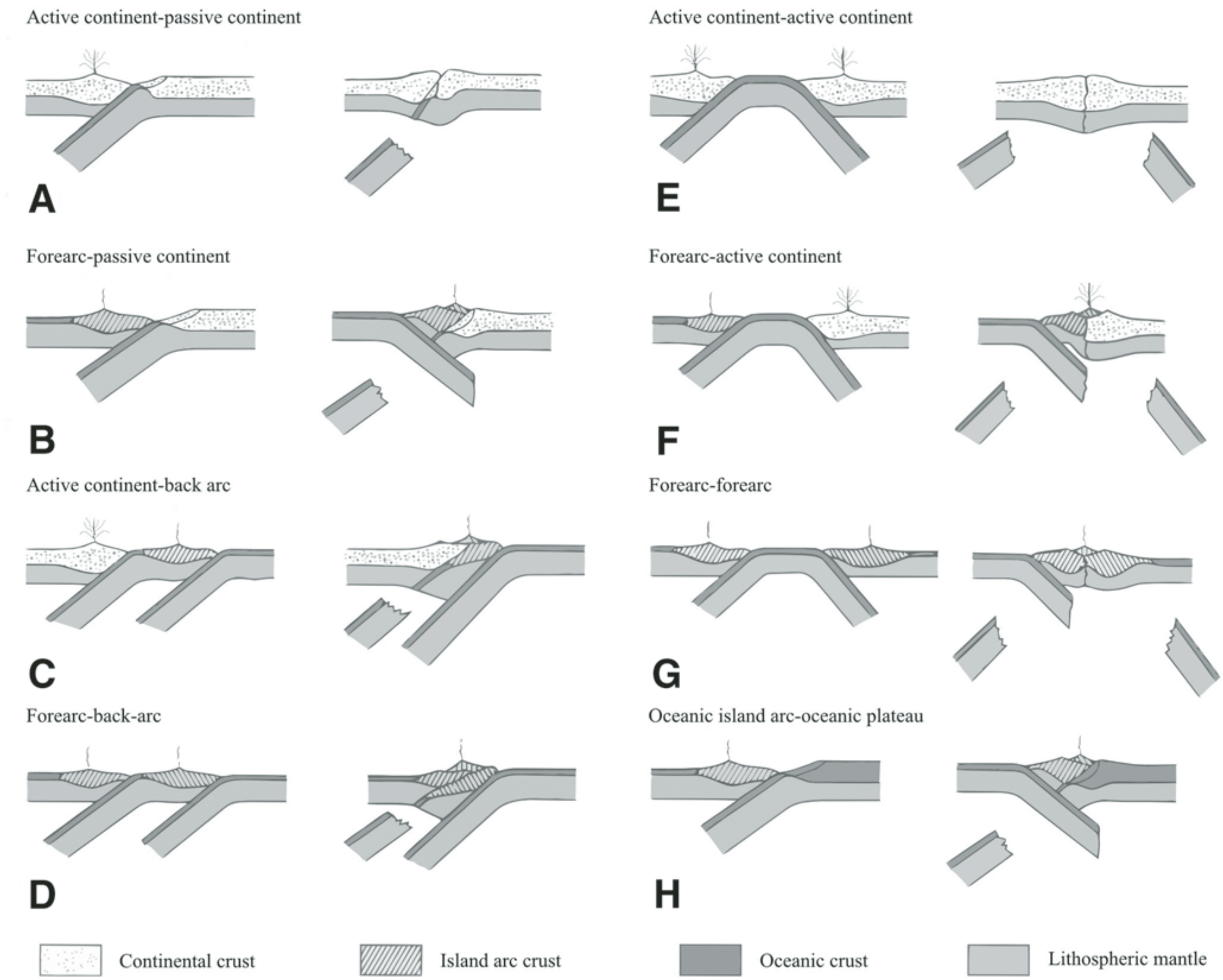


Figure 19. Schematic cross sections of collision types. (A) Active continental margin with passive continental margin, showing collision and breakoff of downgoing slab. (B) Oceanic arc-forearc with a continental margin. Collision results in a "flip" of subduction direction, and breakoff of original downgoing plate. (C) Collision of active continent and backarc of an oceanic arc. One plate breaks off, and the island arc subduction zone remains the active one. (D) Collision involving two oceanic arcs, the backarc of one and the forearc of another. One arc edifice is imbricated under another one, and one subduction zone remains active. (E) Collision between two active continental arcs. After collision a suture forms, subduction ceases, and plate motion must be accommodated elsewhere. (F) Collision between the forearcs of an island arc and an active continental margin. Island arc crust is added to continental margin, old subduction zones cease activity, and a new zone dips under the continent. (G) Collision of the forearcs of two oceanic island arcs. Crust becomes amalgamated, and a new subduction zone could form dipping either to the left or right of the amalgamated arc. (H) Oceanic arc–oceanic plateau collision. Arc and plateau crusts became amalgamated, and a new subduction zone could form dipping under the amalgamated arc-plateau crust. Modified after Twiss and Moores (2007, fig. 19.13, p. 627).

day representatives of the fold-thrust belts known for more than a century in orogenic zones. Thus these modern active regions provide a tie-in with pre–plate tectonic studies of such zones, as discussed below.

The emplacement (or obduction) of many ophiolite complexes involves a collision of an oceanic subduction zone with a passive continental margin or an island arc (see Fig. 20A). In these cases the continent moves into the subduction zone; rocks of the latter are pushed over the continental platform, and fragments of the overriding plate become part of the fold-thrust belt as allochthonous ophiolite complexes. Many examples of such ophiolite complexes in the Alpine-Himalayan belt involve collisions with insufficient subduction to develop an active island arc. Such ophiolites represent so-called *upper plate ophiolites* (Wakabayashi and Dilek, 2003), or that oceanic crust and mantle formed at a spreading center, with development of a subsequent subduction zone, probably along a transform fault or fracture zone, and then were emplaced without sufficient subduction to develop an overlying island arc. Wakabayashi and Dilek (2003) postulate four possible types of ophiolite emplacement, including (1) collision with a continental margin (Tethyan type), (2) collision with an active arc or accretionary complex (Cordilleran type), (3) an igneous intrusion during ridge-subduction zone intersection (rare), and (4) possible exposure of ocean floor during plate motion changes (Macquarie type).

The thrust faults beneath emplaced ophiolites represent remnants of subduction zones. It is perhaps not surprising that many ophiolites display "supra-subduction zone" compositions. Metamorphic rocks ubiquitously occur beneath ophiolite soles (Wakabayashi and Dilek, 2003). The displacements are indeterminate, but they may be hundreds of kilometers. Ophiolites thus emplaced represent major tectonic features, not simply igneous intrusions. Hess's (1939, 1955) recognition of the importance of ultramafic and related rocks in orogenic zones, "in the first stages of orogeny," is still an important statement, but emplacement is a *tectonic* event rather than an *igneous* event.

Ophiolites emplaced as in Figures 20C and 20D are rare. They may represent the result of collision of an oceanic plateau with a subduction zone (type H, Fig. 19). The Malaita ophiolite, southwest Pacific, may represent such an ophiolite formed in this manner (Coffin and Eldholm, 2001).

OROGENIC BELTS

Orogenic (after the Greek words *oros,* meaning mountain, and *genesis,* meaning origin, birth) belts coincide, not surprisingly, with some of the Earth's major mountain belts. Since time immemorial, mountains have held the attention of humans as centers of terror, houses of gods and demons, or sources of inspiration. Geologic exploration of mountain belts, or orogenic belts, began some two centuries ago, and a great deal of information had accumulated prior to the Plate Tectonic Revolution. The formulation of plate tectonics led to a radical revision of ideas on the formation of orogenic belts. This revision has included reinterpretation of old data and collection of new information, utilizing the rapidly developing new ideas and investigative tools mentioned above (cf. Twiss and Moores, 2007, p. 639–692).

Because all oceanic lithosphere in the oceans is less than 200 million years old, the record of any prior plate tectonic activity primarily resides on the continents in orogenic belts. And any

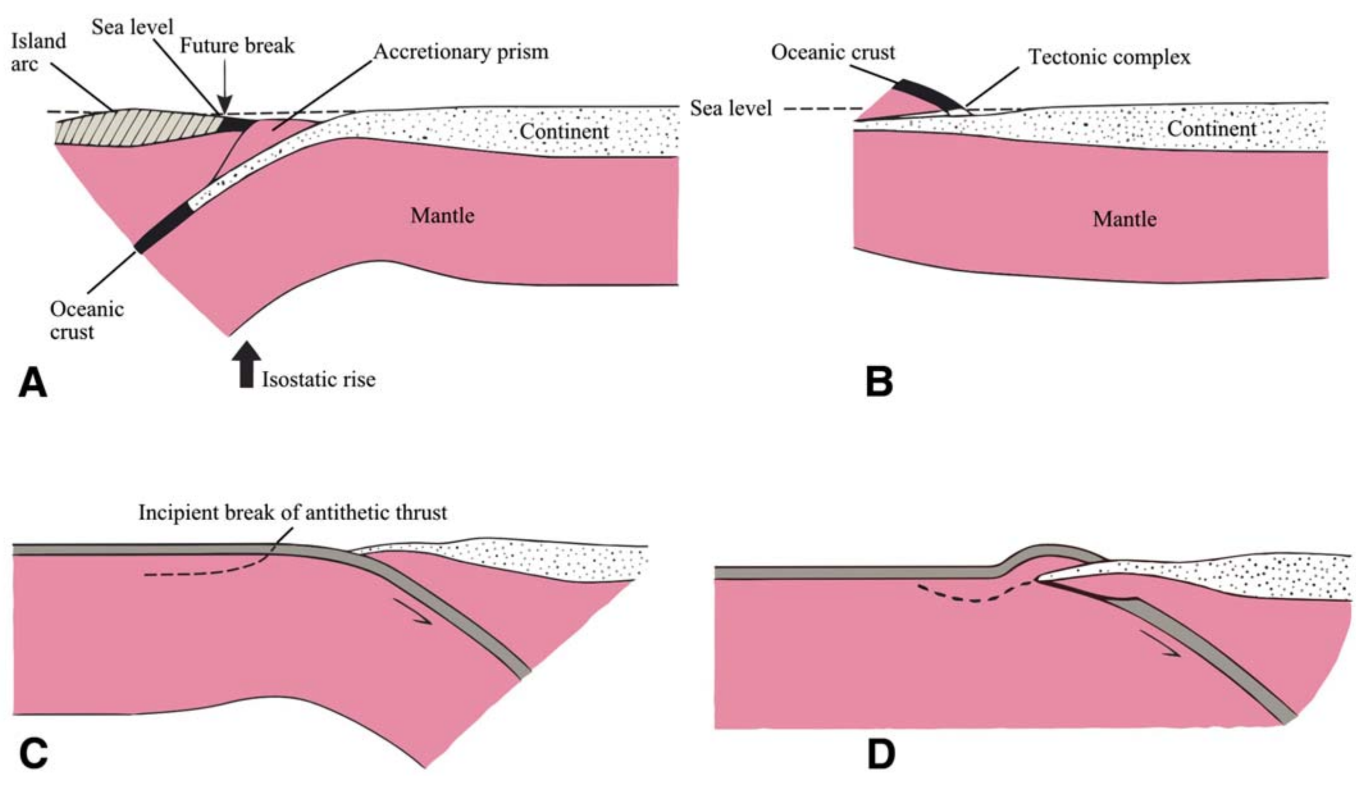


Figure 20. Possible emplacement of an ophiolite. (A) Collision of continent (which could also be an island arc or remnant arc) with a subduction zone dipping away from the continental margin. Underthrusting of crust beneath the mantle eventually chokes off the subduction. (B) Isostatic rise leaves a remnant of the former overriding plate on the continental margin. (C) Development of an antithetic thrust in the downgoing plate. (D) Continental margin wedges under a part of the oceanic plate, emplacing the ophiolite over the continental margin. The scenario in A and B represents the original emplacement mechanism of Moores (1970) and the original obduction of Coleman (1971; see also Dewey, 1976). (Modified after Twiss and Moores (2007, fig. 19.45, p. 629).

information on oceanic crust resides in ophiolite complexes and accretionary prisms.

Orogenic belts range in age from active or Recent, or still active, to Archean. Despite their age, they display some features in common.

There are two principal types of orogenic belt, intracratonic and Cordilleran (Figs. 21A, 21B). Figure 21A shows a model composite cross section of a typical intracratonic two-sided orogenic belt . We interpret such a belt as having formed by collision of two continents; the Himalaya epitomize an ongoing example of such a belt. Such is not the case with all orogenic belts, especially Andean or Cordilleran orogens, along modern plate boundaries or oceanic margins (Fig. 21B; e.g., Ernst, 2005; Dickinson, 2004).

However, the two-sided orogenic belt is common enough to make its description useful. Principal features are as follows, from the margins to the center:

Outer foredeep or foreland basin. Sitting atop the surrounding undeformed continent is a sequence of clastic sediments, up to 8–10 km thick, derived from the rising mountain belt. These sediments coarsen and thicken toward the mountain front, and they commonly display an "unroofing sequence" with clasts derived from progressively deeper levels of the orogenic center.

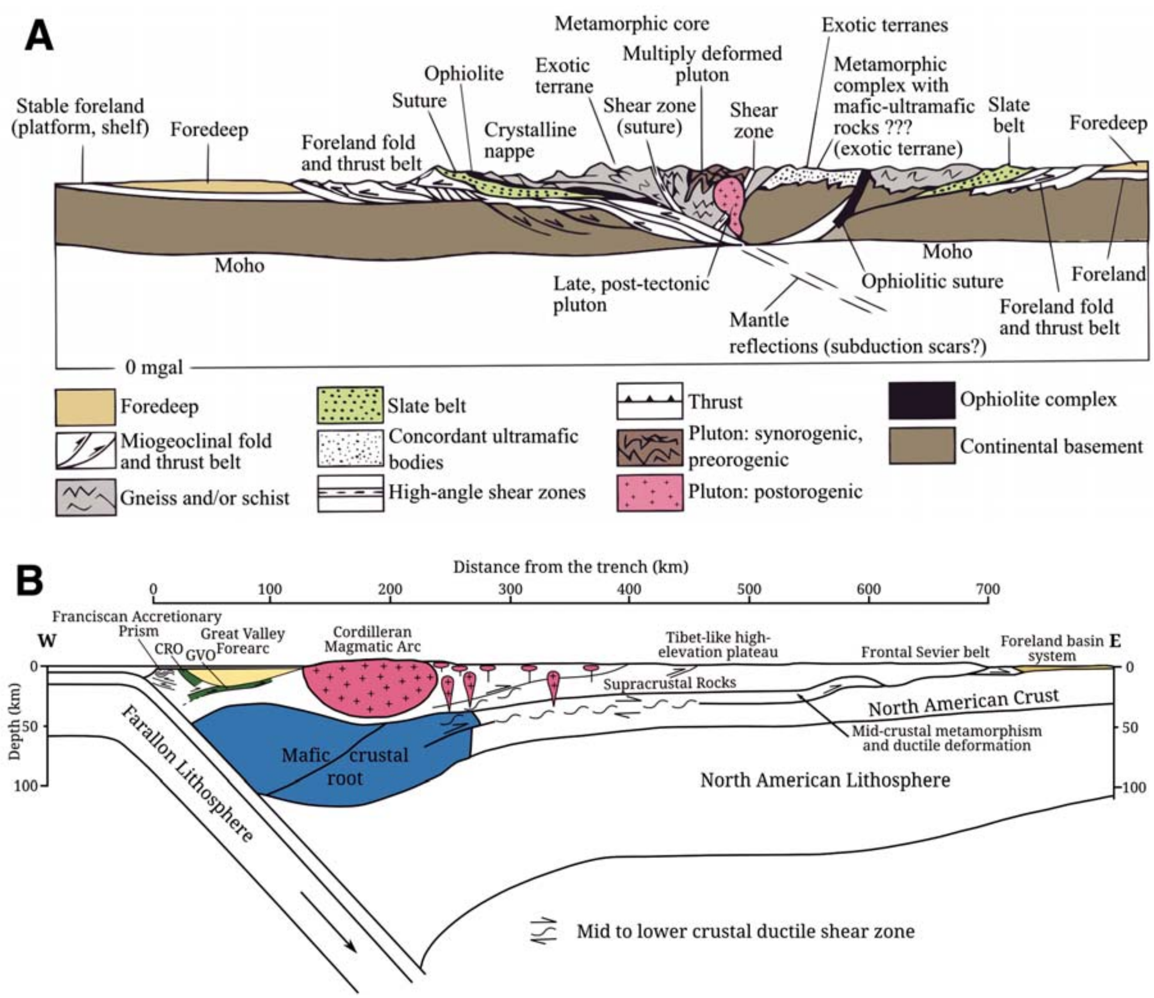


Figure 21. (A) Diagrammatic cross section of a composite orogenic belt, showing formerly separate continental crust on each side, two foreland fold and thrust belts, a composite metamorphic core, possible suture locations, ophiolite zones, sutures both ophiolitic and ophiolite-free, high-angle shear zones, possible exotic terrane, and early multiply folded pluton and late-stage post-tectonic ones; mafic-ultramafic complex of uncertain parentage, a slate belt, mantle reflections possibly reflecting subduction scars. Redrawn after Twiss and Moores (2007, fig. 20.3). See text for discussion. (B) Schematic cross section through a representative Andean-style margin, the Cordilleran margin of the Western United States. Modified after DeCelles and Coogan (2006, fig. 12) and Moores et al. (2002). Cross section represents an ~300 km shortening of crustal rocks along thrust faults. Central Rocky Mountains are present to right of Foreland basin system. Modified after DeCelles and Coogan (2006, fig. 12) and Moores et al. (2002). CRO—Coast Range ophiolite; GVO—Great Valley ophiolite. See text for discussion.

External fold-thrust belt. A fold-thrust belt is directed away from the center of the orogen toward the platform. In most cases it is composed of continental marginal sediments—the "miogeocline" of the geosynclinal view of mountain belts. In its outer reaches the external fold-thrust belt involves rocks of the foredeep, whereas in its internal region (toward the orogenic center), it involves ophiolitic, volcanic, and slightly metamorphosed clastic rocks that are now slates and intercalated foliated sandstones.

External massifs. These are folded, and in some cases metamorphosed, continental basement rocks. In many cases they represent remnants of microcontinents separated from the original continental margin by a narrow ocean basin and swept up in the collision of the two major continents, or else faulted fragments of a rifted continental margin.

Slate belt. Many mountain belts exhibit a monotonous sequence of low-grade metamorphic rocks, wherein continental rise–deep sea sediments have been metamorphosed and thrust toward the platform (e.g., Hsü and Schlanger, 1971). These rocks are generally only sparsely fossiliferous, and correlations are difficult and imprecise. Metamorphism is of high zeolite–lower greenschist facies; multiple folding is fairly common, and axial plane foliation accompanies the folding.

Ophiolites may be present near this position. They crop out predominantly in two ways—as small tectonic slices or blocks of incomplete sequences formed during subduction of an oceanic plate, and complete "upper plate" (Wakabayashi and Dilek, 2003) ophiolitic sequences as large subhorizontal thrust sheets hundreds of kilometers in dimension. Ophiolites commonly overlie a tectonic complex of thrust slices of platform, slope, rise, and abyssal sediments, or a mélange complex (Fig. 22). As mentioned above, the basal contact of an ophiolite thus emplaced represents an example of a fossil subduction zone, i.e., a *suture*—a scar of a disappeared ocean. A given orogenic belt can contain several sutures; their identification and understanding is one of the most challenging problems in deciphering the history of the belt. (e.g., Dewey, 1977, 1987; Moores, 1981).

A suture is a surface or zone separating two non-subductable pieces of crust that are juxtaposed following disappearance of an intervening ocean. Criteria for sutures include the presence of ophiolites (as mentioned above), separate fold-thrust belts, varied or discrepant paleomagnetic directions in rocks of a similar age, a separate metamorphic sequence interpreted as an arc-subduction zone pair, faunal province boundaries, boundaries between radiometric age provinces, fault-bounded rock assemblages with different metamorphic or structural characteristics, and major shear zones containing ophiolitic or other oceanic rock (Burke et al., 1977; Dewey, 1977, 1987; Moores, 1982; Şengör and Natal'in, 1996a).

Metamorphic rocks dominate in the central parts of most orogenic belts. The temperature and pressure conditions at which these rocks have equilibrated have been the product of exhaustive laboratory experimentation. Figure 23 shows some pressure-temperature relations of common metamorphic assemblages. The nature and sequence of metamorphic assemblages reflect the progression of the rock through a temperature-pressure (depth) path. Utilizing the nature of metamorphism and the radiometric ages of metamorphic events, it is possible to get an idea of the history of a rock. In some cases a given rock may exhibit an older high-pressure, low-temperature environment with subsequent metamorphism, with mineral assemblages displaying an intermediate-temperature, medium-pressure environment. In other words, blueschist metamorphism tends to occur early in some rocks, which may be overprinted by greenschist metamorphism (see solid red line with arrows in the figure). Modern analytic techniques enable such effects to be unciphered (e.g., Wakabayashi and Dilek, 2003; Fig. 23). Some orogenic belts show evidence of high temperature, high pressure, and even ultra-high-pressure metamorphism in the field of stability of coesite.

The fields of *Buchan* and *Barrovian* metamorphism are terms derived from the Scottish Highlands, UK (Turner, 1968). Recent work has shown that Barrovian zones may represent overprinting at medium pressures and temperatures of earlier formed high-pressure, low-temperature metamorphism, or conversely, high-temperature, low-pressure metamorphism (e.g., Wakabayashi, 2004; Brown, 1998a). Buchan metamorphism possibly results from subduction of an oceanic ridge (Brown, 1998b, Wakabayashi, 2004). Several metamorphic terranes have yielded evidence of burial to the coesite field of stability (so-called ultra-high pressure, UHP: Carswell and Zhang, 2000; Krabbendam and Dewey, 1998; Liou, 2000).

Belts of regional metamorphism have been described to include so-called *isograd surfaces,* those defined by appearance or disappearance of characteristic minerals. In some cases, careful work can show the three-dimensional nature of these surfaces—roughly parallel to the Earth's surface. In a few cases, e.g., the Himalaya, the mapped isograd surfaces are inverted and may have themselves been folded (Figs. 24, 25).

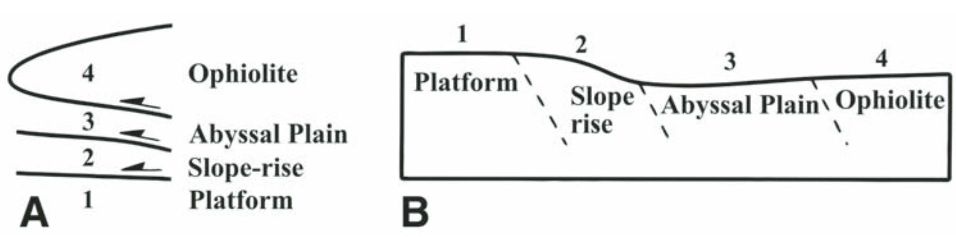


Figure 22. Schematic cross sections of orogenic emplacement of ophiolites. (A) Stack of thrust faults, showing slices of platform, slope-rise, and abyssal plain sediments in successive slices beneath the ophiolite. (B) Thrust slices restored to pre-faulting configuration. Compare with Figure 20. Redrawn after Twiss and Moores (2007, fig. 20.7, p. 48).

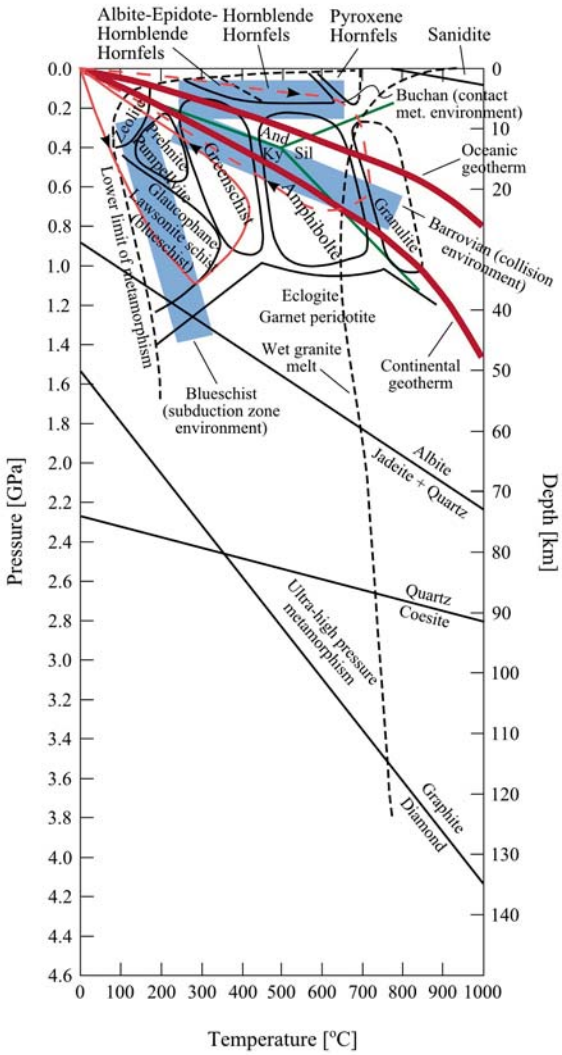


Figure 23. Petrogenetic grid, showing P-T domains of metamorphic facies and critical metamorphic reaction. The blue bands show the approximate locations of metamorphic facies series for conditions of high-pressure, low-temperature (blueschist) metamorphism characteristic of a subduction zone situation; intermediate pressure and temperature (Barrovian) metamorphism typical of "normal" temperature gradients in a collisional situation; and low-pressure, high-temperature (Buchan) metamorphism in a shallow contact metamorphic environment. Shown also is the generalized region of ultra-high-pressure metamorphism below the quartz-coesite transition. Depth scale is approximate and assumes a density of 3100 kg/m^3 (too high for the crust, and too low for the mantle). Heavy red lines show approximate average steady-state geotherms for continental and oceanic crust. Light red line with arrows shows schematic pressure-temperature-time (PTt) path for rock initially subducted into the high-pressure, low-temperature field, undergoing greenschist facies metamorphism, and then exhumed. Dashed line shows schematic PTt path for rock originally metamorphosed in the high T, low P field, then buried in the granulite field, and later exhumed through amphibolite and greenschist fields of metamorphism. Redrawn after Twiss and Moores (2007, fig. 20.8, p. 649).

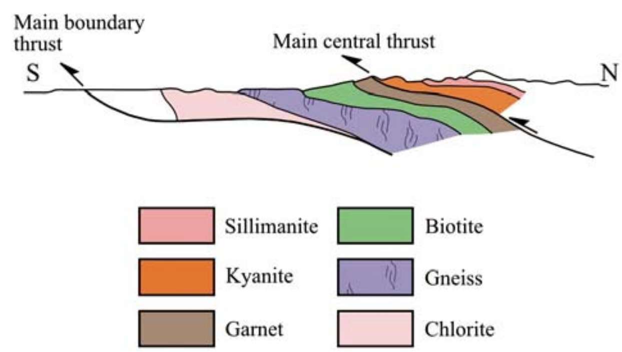


Figure 24. Schematic cross section of the Himalaya, showing metamorphic zones including inversion of isograds. After Le Fort (1975), Gansser (1974), and Moores and Twiss (1995, fig. 10.25B, p. 288).

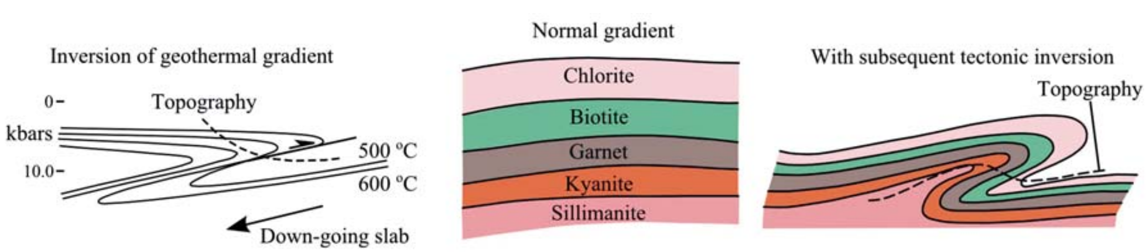


Figure 25. Schematic cross sections, showing possible origin of inverted metamorphic zones by folding of previously formed isograd surfaces. Redrawn after Le Fort (1975) and Moores and Twiss (1995, fig. 10.26, p. 288).

The *crystalline core* zones of mountain belts commonly have metamorphosed rocks that display multiple folding of layers (Fig. 26). The sequence proceeds from isoclinal ductile structures (f_1; Fig. 26A) through secondary open folds (f_2; Fig. 26B) to brittle tertiary kink folds (f_3; Fig. 26C). This sequence of folding may have formed during a progression from highly ductile to more elastic rocks in a collision zone, but the sequence has been known for decades (Ramsay, 1966).

Rocks in the core zone generally comprise one or more of the following sequences (see Moores and Twiss, 1995, p. 274–284; Twiss and Moores, 2007, p. 648–659):

1. Metamorphosed sedimentary rocks and their basement, derived from a continental margin sequence, or from a microcontinent incorporated in the deformation;
2. Metamorphosed volcanic and igneous rocks and associated metasediments, either derived from incorporated volcanic-rich continental margins, oceanic volcanic rocks and associated sediments, a continental or an island arc sequence incorporated in the orogenic belt, or oceanic mid-plate volcanic and associated sediments;
3. Metamorphosed ophiolitic sequences, now present as highly folded rocks or banded amphibolites, deformed mantle peridotites, and associated serpentinites;
4. Lower continental crust and mantle, displaying highly deformed and retro-metamorphosed gneissic rocks and peridotites;
5. Gneissic terranes with abundant ultramafic bodies that may represent metamorphosed rocks from a volcanic-rich continental margin or from deformed and metamorphosed mélange complexes;
6. Granitic batholiths—large masses of igneous rocks of granitic composition, commonly either of I type, formed by partial melting of a hydrous mantle or previously crystallized igneous rocks, or S type, formed by partial melting of sedimentary or metasedimentary rocks. These magmas commonly intrude upward into already deformed rocks, some exhibiting sutures (e.g., Dickinson, 1970b, 2004; Glazner et al., 2004; Hamilton and Myers, 1967; Şengör and Natal'in (1996b). Batholiths may be subhorizontal in shape (Hamilton and Myers, 1967), and they may be folded (Buddington, 1959; Hildebrand, 2013, and references therein).

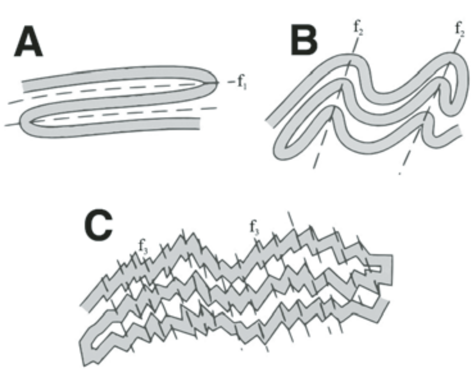


Figure 26. Diagrammatic cross sections, illustrating multiple sequence of folding proceeding from (A) first-generation isoclinal folds to (B) more upright second-generation folds superposed on the earlier deformation to (C) a third-generation kinking superposed on all earlier foldings. After Ramsay (1966) and Twiss and Moores (2007, fig. 20.13, p. 653).

Fabrics in metamorphic regions of mountain belts have long been a source of study (e.g., Krabbendam and Dewey, 1998; Platt, 1986). These fabrics give some idea of transport directions during deformation. These analyses give sense of shear ranging from parallel to perpendicular to the main strike of the mountain belt. In addition, analysis of multiple fold transport directions in several orogens indicates a shearing parallel to the strike of the orogen (Hansen et al., 1967).

Most mountain belts display a *root,* an area where the Moho is depressed to 60 km or so, thus substantially thicker than the 35–40-km-thick normal continental crust. In most orogenic belts the formation of this root is tectonic by formation of large-scale thrust complexes or nappes and refolding of them. In some regions, for example, in the Alps, large, subhorizontal thrust sheets or nappes have been shoved in one direction, and then backfolded into another direction (e.g., Pfiffner et al., 2000; (Fig. 27). Interleaving of ophiolites and crystalline nappes in the figure points toward an original paleogeography of small ocean basins and intervening microcontinents (e.g., Trümpy, 1960; Smith, 1971; Dewey et al., 1973).

One plate tectonic model for formation of such a root incorporates subduction in one direction, continental collision, and then subduction in the other direction to form an Alpine back (root) fold. In outcrops, two episodes of folding might be observed that reflect this succession of different subduction directions (Fig. 28). This model is supported by offsets in Moho discontinuities in the Alps in particular (e.g., Waldhauser et al., 1998), and in other orogenic belts, especially in Tibet (e.g., Xu et al., 2010; Zhu and Helmberger, 1998) and some Precambrian regions (Hammer et al., 2011).

High-angle fault zones are another ubiquitous feature of orogenic belts, also displayed schematically in Figure 21A. These faults generally form late in the history of a belt, by movement of material along strike as it compresses, perhaps owing to a adjustment of material along strike of the two colliding continental margins or from oblique subduction. In some cases movement of material occurs perpendicular to the main movement of plates (e.g., Molnar and Dayem, 2010; Dewey and Şengör, 1979). The large-scale intracontinental strike-slip faults described above would fit this category.

The Himalaya-Tibet region is an example of an ongoing continent-continent collision. (Fig. 29A). India has collided with Tibet, throwing up the Himalaya (e.g., Le Fort, 1996). This complex collision may involve a collision of a passive continental margin with an Andean-style orogenic belt (the Karakorum;

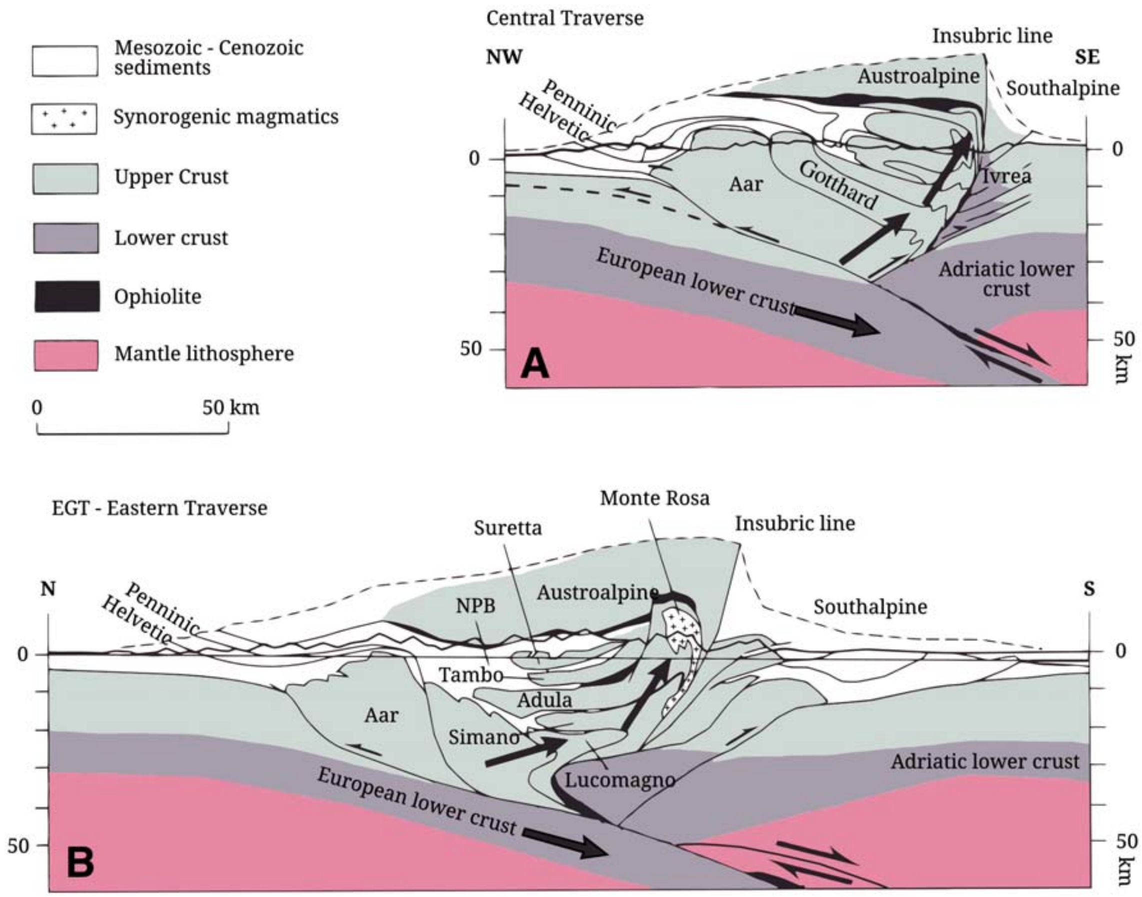


Figure 27. Crustal-scale cross sections of the Alps, showing formation of a mountain root by backfolding of nappe structures. Bold solid line is present topography. Helvetic: nappes of former European continental margin. Penninic: nappes of deeper water metamorphosed clastic rocks including ophiolites. Aar, Gotthard, Simano, Lucomagno, Tambo, and Adula are separate crystalline, multiply folded nappes separated by thin ophiolitic tectonic slivers. Ivrea-Insubric line represents main boundary between Adriatic and European rocks. European lower crust and mantle dip beneath Adriatic lower crust and mantle. Heavy arrows—direction of tectonic movement. Austroalpine—displaced piece of Adriatic crustal rocks. (A) Central traverse between Italy and central Switzerland. (B) Between NE Italy and Austria. Redrawn after Pfiffner et al. (2000) and Twiss and Moores (2007, fig. 20.20, p. 661).

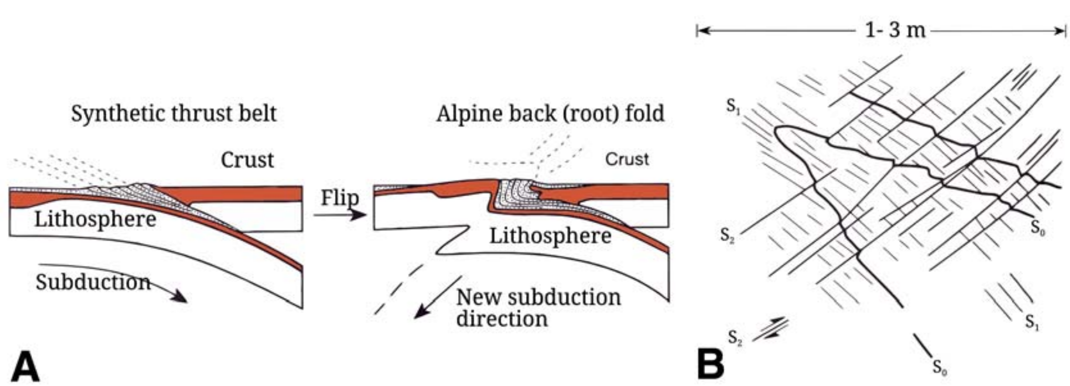


Figure 28. Plate tectonic model for formation of an Alpine root zone by collision of two continents and flip of the subduction direction. After Roeder (1973) and Twiss and Moores (2007, fig. 20.21).

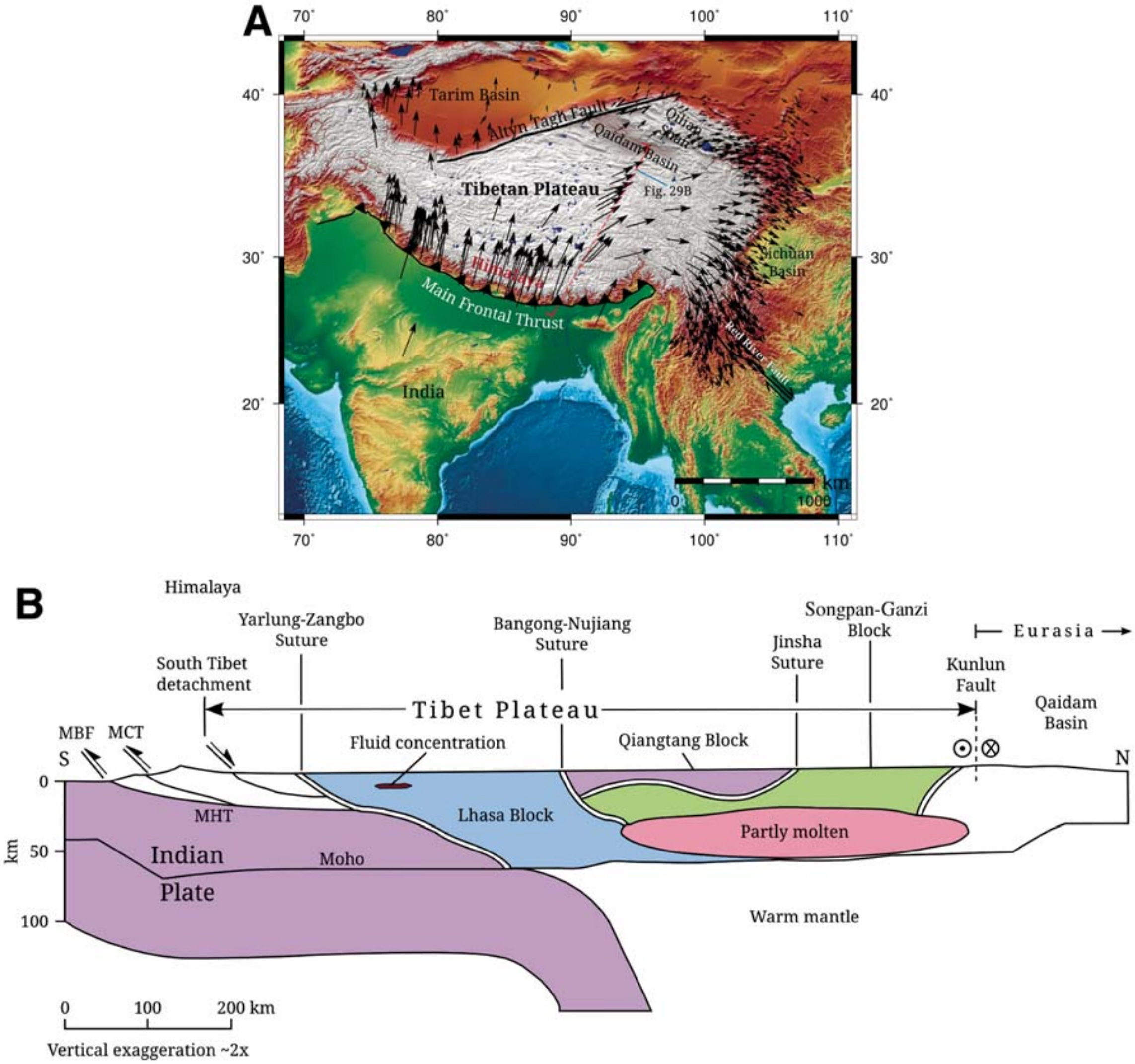


Figure 29. (A) Digital elevation map and GPS motion vectors of the Tibetan Plateau and surrounding region (Amante and Eakins, 2009; Wessel and Smith, 1991, 1998). Redrawn after Twiss and Moores (2007, fig. 6.2, p. 136) and Tapponnier and Molnar (1976, p. 320). Dashed red line is approximate location of Figure 29B. Black arrows are GPS vectors after Zhang et al. (2004). (B) Generalized cross section of the Himalaya-Tibet region (see Fig. 29A for location). The Indian plate (violet) dips northward beneath Tibet, and the Himalaya are composed of rocks from the now-subducted passive margin formerly along the northern margin of the Indian continent. The Lhasa, Qiangtang, and Songpan-Ganzi blocks are microcontinents or intraoceanic island arcs caught up in the Indian-Eurasian collision. The Bangong-Nujiang and Jinsha sutures are of Middle Jurassic age, and the Yarlung-Zangbo suture is of Tertiary age and separates Indian rocks from those of the terranes to the north. The Main Boundary fault (MBF) is the main lower active thrust fault of the Himalaya; the Main Central Thrust (MCT) is a major thrust fault emplacing metamorphosed rocks over unmetamorphosed ones. They coalesce in the Main Himalayan Thrust (MHT). The South Tibetan detachment zone is a north-dipping detachment (normal) fault active during southward thrusting and uplift of the Himalaya. All sutures contain ophiolites. Modified after Haines et al. (2003).

Searle, 1996) with intervening arcs (Haines et al., 2003). Even as India continues to move northward, the Himalaya are also being reduced in elevation by detachment faulting that proceeds simultaneously with southward thrusting (e.g., Burchfiel et al., 1992; Kellett and Djordje, 2012; also in this case, movement of material is eastward and around the bend or *syntaxis*. One interpretation is that there is a so-called *channel flow* in middle crustal levels of rock toward the east to make room for the Indian subcontinent as it continues to move northward toward Asia (e.g., Thatcher, 2009; Clark and Royden, 2000). The Tibetan Plateau apparently ends against the Longmenshan fold and thrust belt on the western side of the Sichuan basin in western China (Dong Jia et al., 2006; Kirby et al., 2002). The 2008 Sichuan earthquake occurred along this belt.

Figure 29B shows a north-south cross section of the Himalaya-Tibet region, crossing in eastern Tibet. The figure shows that the Indian plate dips northward beneath Tibet. The Main Boundary fault is the principal edge of the Himalaya proper. The South Tibetan detachment is a north-dipping fault that is active along with compressional movement, and the Main Himalayan thrust connects with the Main Boundary thrust. The Lhasa, Qiangtang, and Songpan-Ganzi blocks are microcontinents or intraoceanic island arcs caught up in the Indian-Asian collision and separated from each other by ophiolite-bearing sutures. The Qaidam Basin is a basin near the Tarim continental block, which is part of the Asian continent. Seismic evidence indicates that fluid concentrations are in the upper crust of the Lhasa block, a partly molten region beneath northern Tibet. The lower crust beneath eastern Tibet is partly molten, and the mantle below Tibet is also warm. Both the mantle and the crust may be moving by ductile flow toward the east (Haines et al., 2003). This evidence for mid-crustal flow of ductile rocks parallel to the trend of the orogenic belt recalls the textures indicating such a flow in ancient orogens as mentioned above.

Continental marginal (Cordilleran) orogenic belts such as the example displayed in Figure 21B are two-sided as well (e.g., Burchfiel and Davis, 1972, 1975; DeCelles and Coogan, 2006; DeCelles, 2004). However, the two-sidedness of these belts is primarily in the presence of thrusts and folds in two directions. In the North American Cordillera, exemplified by the cross section in Figure 21B, the fold-thrust belt on the east side of the mountain belt consists of a set of west-dipping, east-vergent structures—the so-called Sevier fold-thrust belt that has moved toward the North American craton. Basement-cored faulted uplifts of the Central Rocky Mountains lie east of the section. The other "side" of the orogenic belt represents structures in the California Coast Ranges and elsewhere that parallel the current or past subduction zone; that is, they dip generally eastward and are primarily west-vergent. The plate west of the subduction zone is the downgoing oceanic plate, or along the transform margins, the Pacific plate.

The North American Cordillera is a long-active system, with activity extending back into the late Precambrian (e.g., Moores et al., 2002). External massifs are present in the form of the basement-cored faulted and uplifted Central Rocky Mountains. A complex, multiple-aged slate belt is present in the form of deformed early Paleozoic abyssal shales in central Nevada, an extensive belt of slaty and phyllitic multiply deformed metamorphic rocks in the Sierra Nevada ranging in age from pre-Ordovician to Cretaceous. Ophiolite-bearing sutures are widespread (e.g., Ingersoll, 1998; Ingersoll and Schweickert, 1986; Moores et al., 2002). An extensive granitic batholith crops out along much of the length of the belt in the Peninsular Ranges of southern California and northern Baja California, and the Sierra Nevada of California and Nevada, Idaho, and British Columbia. The metamorphic grade of the batholith wall rock is regionally low, and high-temperature, low- to medium-pressure rocks are locally present (e.g., Day et al., 1988; Hacker, 1993). Metamorphic core complexes are widespread but are separated from each other by less recrystallized rocks (e.g., Armstrong, 1982). Volcanic rocks are abundant.

A key difference between Cordilleran-type orogens and those of the more "classic two-sided orogeny," as outlined above, is the possible cause of deformation. The conventional hypothesis is that Cordilleran orogens are substantially noncollisional, and that deformation in the fold and thrust belt is antithetic to subduction and is caused by flattening of the subducting slab, by pressure from the subducting slab, or by counterflow and accumulation of lower crustal and mantle material against the subduction zone. The fold and thrust belt is antithetic to a single subduction zone active during the orogeny (e.g., Dickinson, 2004; Ernst, 2005; Ducea, 2001). A countervailing view is that Cordilleran orogens are in fact collisional, just as with the intracratonic orogens (e.g., Chamberlain and Lambert, 1985; Hildebrand, 2013, 2009; Johnston, 2008; Lambert and Chamberlain, 1988; Mattauer et al., 1983; Moores et al., 2002; Moores, and Day, 1984; Moores, 1970, 1998), and that the foreland fold and thrust belt is synthetic (i.e., parallel) to the subduction direction of colliding island arcs or continental slivers (ribbon continents) that were emplaced along subduction zones dipping away from the continent. Large-scale orogen-parallel movements are also present (e.g., Hildebrand, 2013, and references therein).

Perhaps Cordilleran-type orogens are not really distinct from intracratonic belts; rather they represent an intermediate stage in continental reassembly, similar to what happened in central-eastern Asia (Şengör and Natal'in, 1996b), wherein a complex series of island arc assemblages and continental fragments amalgamated over hundreds of millions of years to produce the very large Asia-India continent. The presence of an Andean margin in the Karakorum (Searle, 1996) and at other times in other parts of Asia (Şengör and Natal'in, 1996b) suggests that such a process has been involved in development of the Himalaya-Tibet and neighboring regions. A similar collisional model for the modern Andes is also possible (e.g., Moores et al., 2002; Hildebrand, 2013).

In conclusion, orogenic belts comprise present and past great mountain belts of the Earth. Most belts reflect continent-continent or continent-island arc collisions. Many structures in orogenic belts are subhorizontal, and most belts are two sided, with thrusts over forelands on both sides. Orogenic belts form during assembly and breakup of supercontinents, but most continents do not converge and collide with each other in the same locations that they rifted way from. Modern tectonic activity in the western Pacific (Hall, 2002; Schellart et al., 2006) illustrates intraoceanic complexity that may be assembled between two collided continents. Circum-Pacific deformed belts represent partially developed orogenic belts, pending the ultimate closure of the Pacific Ocean.

In convergence and collision, continents converging along subduction zones sweep before them and collect intervening island arcs, other oceanic islands, and microcontinents. Orogenic belts exhibit ophiolites (oceanic crust preserved by early continent–reverse-dip subduction zone collisions), remnants of oceanic collisions, rifts, subduction zone flips, multiple sutures, intraoceanic strike-slip faults, pre- and synorogenic granitic intrusions and arc deposits. Collided pieces adjust to continental margin shapes by local strike-slip faults and thrusts of possible opposite sense. Major subhorizontal thrust complexes cut all rocks. Mountain roots and major topographic welts form and are worn down by chemical changes and mantle-crust adjustment, exhumation, and erosion, respectively. Canadian Shield observations (Hammer et al., 2011) indicate that similar processes have operated since the Neoarchean (2.7 Ga). Magmatic oceanic crust sections were thicker prior to 1 Ga. Ophiolites are present in pre–1 Ga orogenic belts as thrust slices of mostly magmatic rocks (greenstone belts) with only rare mantle exposures (Moores, 2002).

TECTONICS, TOPOGRAPHY, AND EROSION

Since development of the plate tectonic theory, it has become clear that tectonic forces in collisions are major factors in vertical uplift associated with mountain belts. Three distinct but related concepts are involved with uplift: *Surface uplift* is the vertical displacement of the Earth's surface with respect to the geoid; *rock uplift* is the vertical displacement of a given volume of rock with respect to the geoid; and *exhumation* is the vertical displacement of a given volume of rock with respect to the topographic surface. Thus, exhumation is the difference between rock and surface uplift, i.e.,

exhumation = rock uplift – surface uplift,

and similarly,

the rate of exhumation = the rate of rock uplift –
the rate of surface uplift.

For a given volume of rock, if the surface uplift equals the rock uplift, the rock volume does not approach the surface, and the exhumation is zero.

Rock and surface uplift can occur together as a result of isostatic uplift or tectonic crustal thickening. Processes that produce such thickening include offscraping and underplating at a subduction, horizontal shortening and vertical thickening of part of the crust by thrusting or folding, delamination of dense lower crust and its replacement with hot buoyant asthenospheric mantle, and rifting of continents, thinning the lithosphere. In such cases, exhumation occurs by removal of material by erosion or by tectonic denudation along low-angle normal faults that in effect lengthen and thin the crust. Thus when

exhumation = erosion + tectonic denudation,

then

rock uplift – surface uplift = erosion + tectonic denudation,

and the rock uplift rate must exceed the surface uplift rate. Rock uplift without surface uplift can be present in such settings as diapiric rise of a less dense rock through a denser one, or in return flow along a subduction zone.

Generally high topography is supported isostatically by thickened low-density crust or by lower mantle density directly beneath the crust. Isostatic surface uplift can occur by thickening of the crust during orogenic shortening or by replacement of high density mantle by low density mantle. Such replacement can occur two ways, by extension of preexisting crust and mantle through a rifting and tectonic thinning of the crust, and by *delamination,* separation of high density mantle from overlying crust and its replacement by low density asthenospheric mantle.

There is an intimate relationship between tectonic uplift and erosion. Erosion rates and rock uplift rates for orogenic belts are about equal, ranging from approximately tenths of a millimeter per year to several millimeters per year (1 mm/a = 1 km/Ma). High-grade metamorphic rocks that come from 20 to 40 km depth are present at the surface of many mountain ranges. In some places, ultra-high-pressure metamorphic minerals are found, indicating that the rocks have gone down to greater than 100 km depth and then returned to the surface. Equivalent thicknesses of rocks have to be eroded in some way in order for such rocks to be present. To produce high mountains, such as the Alps or the Himalaya, uplift rates exceed erosion rates. However, as the forces of uplift diminish, the topographic elevation gradually declines to a small amount, such as with the Appalachian Mountains of eastern North America, or the Pan-African–Braziliano ranges of former Gondwana.

Understanding the relationship between tectonics and topography is making rapid progress. New means of dating erosional surfaces enable the measurement of long-term uplift and erosion rates. New surface motion measurement techniques, involving the global positioning system (GPS), interferometric synthetic aperture radar (InSAR), and differential light detection and ranging (LIDAR) (Oskin et al., 2012) enable the measurement of uplift and/or horizontal motion rates.

The relationship between tectonics, topography, and climate has also been a source of vigorous debate (e.g., Ruddiman, 1997; Molnar, 2009, 2003; Molnar and England, 1990; Searle, 1996). Items of discussion have included whether the south Asian monsoon has resulted from the uplift of the Himalaya (Ruddiman, 1997; Searle, 1996), and the climatic effect of weathering and resultant sequestering of CO_2 (e.g., Kirchner et al., 2001; Riebe et al., 2004, 2001). These authors effectively showed that tectonics has an important influence on erosion rates, and that chemical weathering rates depended on the latter more than climate.

GEOCHRONOLOGIC METHODS IN TECTONICS

Many advances in geochronology have helped refine ideas of tectonics (e.g., Parrish et al., 1988; Day and Bickford, 2004; Bickford et al., 2006; and many others). These advances include refinements in existing methods, e.g., ^{40}A-^{39}A, U-Pb, and development of powerful new instruments. New methods including Lu-Hf, Os-Ir, and refined fission-track methods, are having a huge effect on the interpretation of tectonics.

These radiometric methods are useful for determining the age of intrusive or extrusive rocks or to gain an understanding of the nature of sources of magmas. In some cases, radiometric techniques can be used to date activity on faults. It is also now possible to collect multiple Pb-U dates on multiple detrital grains of zircon in sediments in orogens. This possibility has opened up a new field of understanding the provenance of the sediments (e.g., Bernet and Spiegel, 2004; LaMaskin, 2012; Wright and Wyld, 2006).

Cosmogenic nuclides, particularly ^{10}Be, ^{26}Al, ^{36}Cl, ^{14}C, ^{3}He, ^{21}Ne, and ^{38}Ar, can now be used to date erosional surfaces. The two most frequently measured cosmogenic nuclides, ^{10}Be and ^{26}Al, are formed by cosmic ray spallation of oxygen and silica in quartz. Another important isotope, ^{36}Cl, is produced by spallation of Ca or K. Depending on rock and landform weathering rates, minimum ages are obtainable for landforms ranging in age from a few hundred years to tens of millions of years (Ivy-Ochs and Kober, 2007; Schaefer and Lifton, 2007).

"WILSON CYCLE"

Wilson (1968) proposed a six-stage orogenic cycle of continent rifting, ocean opening, and then convergence and collision. His steps included (1) embryonic (as in East Africa), (2) young (Red Sea, Gulf of Aden), (3) mature (Atlantic Ocean), (4) declining (western Pacific Ocean), (5) terminal (Mediterranean Sea), and (6) relict scar (geosuture; Indus suture, Himalaya). As ideas on plate tectonics and orogenic evolution have developed, it is possible to propose an expanded "Wilson cycle" (Burke, 2011). In calling these steps a cycle, and in working with schematic two-dimensional cross sections, Wilson was apparently thinking of continents rifting apart, forming a mature ocean basin, and then coming together again more or less in the same position. We now know such has not always been the case. Continents rift apart, forming ocean basins, then converge and collide, but there is not really anything cyclic about this process. There is no guarantee that two continents will converge and collide along the same margin from which they rifted; indeed, there is no guarantee that these continents ever were previously in contact with each other. The history of the assembly and breakup of either Pangaea or the earlier supercontinent, Rodinia (e.g., Li et al., 2008), indicates that the process is likely less cyclic than a quasi-random walk of various continental fragments and associated island arcs with times of maximum and minimum assembly.

Nevertheless, it is possible to visualize a modified "cycle" or orogenic progression, as follows (see Fig. 30):

- A, B. Rifting of a continent and opening of a new ocean basin, gradually producing a passive margin on both sides of the ocean (Figs. 30A, 30B).
- C. Development of subduction zones within the ocean basin, possibly by conversion of transform fault–fracture zones to subduction zones through change in pole of rotation or relative motion of the plates; or by migration into the ocean of subduction zones from outside ("infection": Mueller and Phillips, 1991; Fig. 30C).
- D. Collision of continental margins with subduction zones, emplacing ophiolites, and causing a first deformation (Fig. 30D).
- E. Production of Andean continental margin and further convergence (Fig. 30E).
- F. Arrival of second continental margin at the subduction zone, producing a continent-continent collision (Fig. 30F).
- G. Adjustment of continental margin collision zone by strike-slip movement of pieces sideways to solve the "room problem"(Fig. 30G).

Sutures would have formed in stages C, D, and F, and would have been modified in stages E and G.

Wilson cycles thus described probably characterize orogenic belts from Archean to Recent (Hammer et al., 2011). This fact implies that some type of plate tectonic activity has been going on since Archean times. As mentioned above, continents do not usually converge and collide in the same places where they rifted. Thus there is considerable need for along-strike adjustment in a colliding belt. Also, oceanic island arcs and microcontinents are invariably involved in the amalgamation of a continental region (Şengör and Natal'in, 1996a, 1996b).

TERRANE ANALYSIS

Many, if not most, of the world's orogenic belts include a composite of distinct segments that originate not only from the continents or arcs involved in the final collision but also from regions that are clearly "exotic" to the main crustal blocks, or whose origin is unknown or "suspect." These segments have become known as *terranes,* and a form of analysis has evolved for understanding these regions. Terrane analysis involves careful separation of contiguous regions from those that were separate, dating, from stratigraphic or radiometric means, the timing of collision (or docking) of such terranes. In most cases the boundaries between such terranes represent sutures or strike-slip faults modified by shortening events.

Terranes represent multiple features:

- Composite arcs. Multiple arc complexes have had a long oceanic history and have undergone collisions in an oceanic region before colliding with a continent. For example, complex assemblages of oceanic arcs and continent-derived clastic sediments are present in the Stikinia-Sierra terranes, Western North America. Some of these arcs display an oceanic history of tens of millions of years. Analysis of zircons in continent-derived sediments suggests that the latter were derived from the other side of the North American continent, or even Africa or Baltica (e.g., Moores, 1998; Moores et al., 2002; Wright and Wyld, 2006). The assembly of Asia was a long, complex amalgamation of separate continental blocks and island arcs (Şengör and Natal'in, 1996a, 1996b).

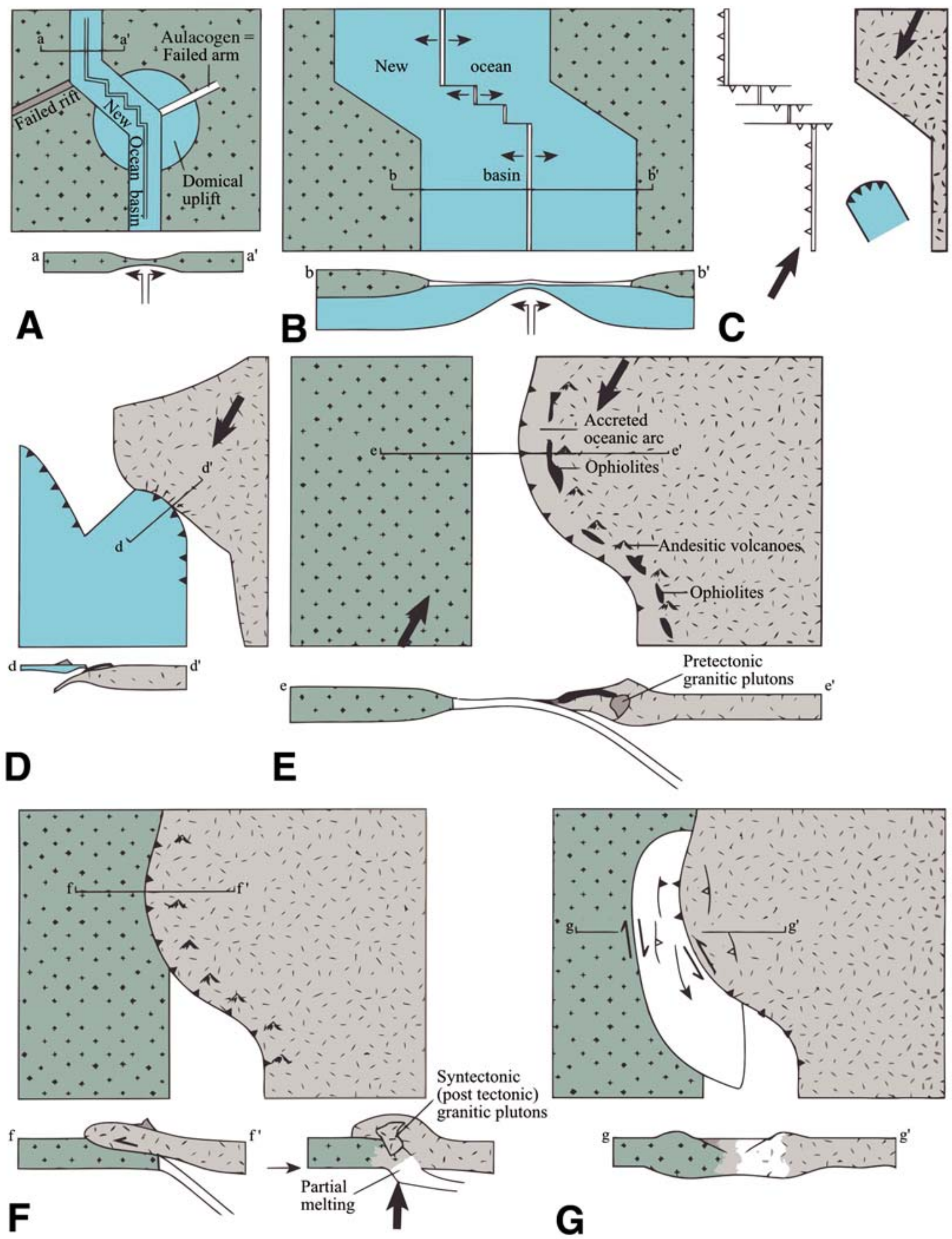


Figure 30. Expanded "Wilson cycle" of orogenic development. Diagrammatic maps and cross sections, showing possible successive plate tectonic stages of rifting and subsequent collision of continents to account for traditional observations of mountain belts. (A) Rifting of continental margin, and formation of domical uplifts and aulacogens across failed arms. (B) Development of mature Atlantic-style ocean with passive margins. (C) Change of ocean from opening to beginning of closing by change in relative plate motion, formation of thrust faults (incipient subduction zones) on ridge-parallel faults, transform faults, and fracture zones. Diagram might also include entry of oceanic arc with its subduction zone into ocean, the "infection" of Mueller and Phillips (1991). (D) Collision of intraoceanic subduction zone with continental margin, and emplacing an ophiolite. (E) Reversal of subduction direction on collided continental margin, and formation of Andean continental margin on previously deformed, ophiolite-bearing continental margin. (F) Collision of two continents with Andean- and Atlantic-style continental margins; formation of mountain root. (G) Adjustment of collision zone by strike-slip movement of pieces, or renewed rifting to "smooth out" the collision zone. Modified after Moores and Twiss (1995, fig. 10.35, p. 298).

- Ribbon continents. These are continental fragments that are much longer than they are wide. They contain crystalline continental crust and overlying supracrustal rocks that reflect a continent's complex history. Some ribbon continents may originally have been more equant in shape than at present. Such highly elongate continental fragments may have originally been a series of separate fragments, or a more equant continental or island arc assemblage that collided and became "smeared out" during continued convergence between two continents on either side of the orogeny (e.g., Johnston, 2008; Hildebrand, 2009, 2013; Şengör, 1979, 1984).
- Individual arcs or island arcs with relatively simple histories, such as the Lesser Antilles, the South Scotia Arc, or the Aleutians, may have been present in an ocean basin, and during closure of that basin migrated toward and collided with a continental margin.
- Ophiolites. As outlined above, ophiolites are fragments of ocean crust and mantle formed at oceanic spreading centers and preserved on continents. Most ophiolite exposures represent partial sections. Individual belts may extend for a few hundred to a few thousand kilometers in length, but complete sections are relatively rare.
- Stratigraphic sequences with distinct lithologies and faunas. Interpretation of lithologies is fairly straightforward. Faunal comparisons, however, are often problematic. A comparison of shelf benthonic invertebrates in the modern Indo-Pacific Ocean indicates that they are essentially similar for >10,000 km, from east Africa to the Tuamotu Islands (Valentine and Moores, 1974). Thus faunal similarity does not necessarily reflect close proximity of original sites of deposition.

As mentioned above, the recognition of exotic terranes in mountain belts has necessitated a major modification of the Wilson cycle. Convergence and collision of two separate continents likely involves a complex array of multiple subduction complexes and oceanic terranes that collided with each other and/or a continent. Before the final continental collision, the collision of exotic terranes with a continent will produce evidence, in the form of a deformed continental margin or deformed terranes themselves, of "docking" of the latter against the former. The development of any orogenic region thus includes the history of the continents themselves, and any deposits that lie within the ocean separating two continents before they finally came together. A given orogenic belt thus includes many episodic events, which taken together, represent a quasi-continual process, with major collisions having occurred only episodically.

Finally, a cautionary note: In order for terrane analysis to be useful, it must carefully separate rocks of different tectonic origins. Simply labeling a large region of metamorphic and igneous rock as the "X" terrane does not help understanding. For example, the so-called Northern Sierra Terrane in the northern Sierra Nevada, California, may have as many as seven sutures (Moores, 2009). To call such a diverse area a single "terrane" is no advance over pre–plate tectonic fixist geological analysis.

HISTORICAL ANALYSIS

With the background outlined above, it is interesting to change focus slightly, and move to specific analyses and speculations on the history of the Earth. We begin with speculative reconstructions of the Earth's interior, then proceed to possible links between large magmatic provinces and patterns of continental assembly and fragmentation.

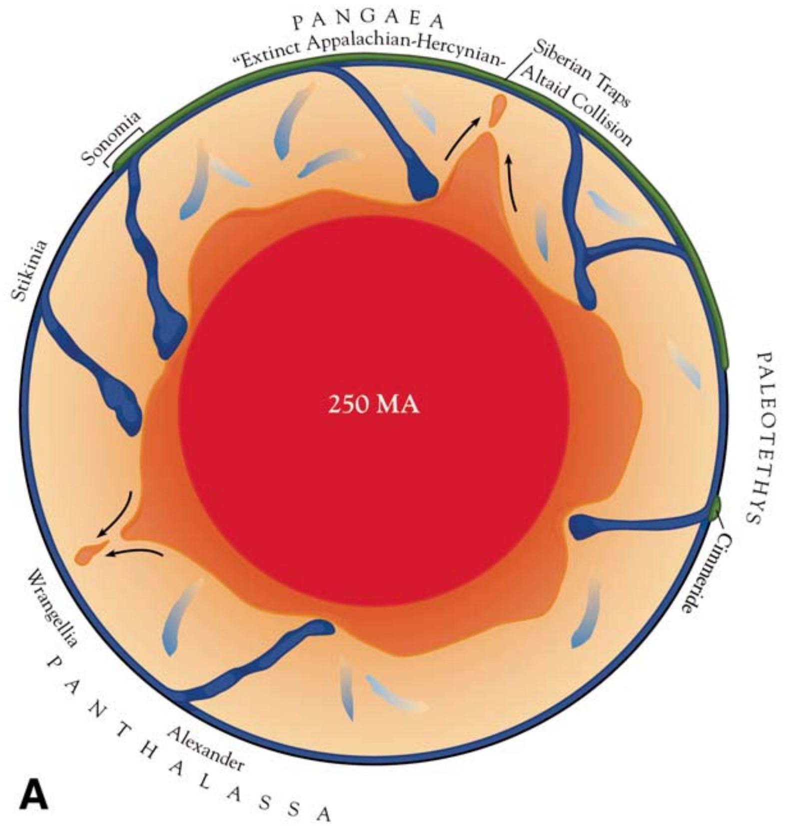
Speculative Reconstructions of the Earth's Surface and Interior

Using the model of the Earth's interior presented in Figure 3, we have inferred the structure of the Earth's mantle at the present time. Here we use the same model of the Earth's interior to speculate on possible crust-mantle structure and history of the Earth in times past. Schematic cross sections (Figs. 31A–31C) through the Earth at 250 Ma, 150–190 Ma, and 100 Ma are an attempt to draw possible cross sections of the Earth showing the structure of the mantle at these specific times (see Moores et al., 2000).

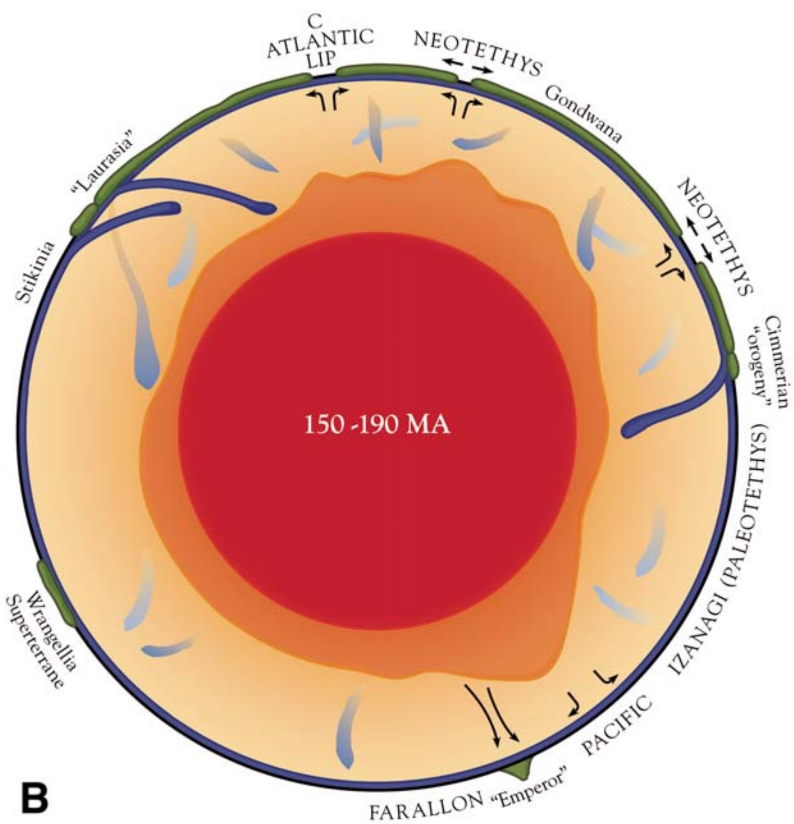
Figure 31A shows a schematic cross section of the Earth at ca. 250 Ma. Pangaea had formed by the just-completed Appalachian–Hercynian–Ural-Altaid collisional orogens between Laurasia (North America) and Gondwana, and Ural (Siberia and Laurasia) and Altaid (island arcs and terranes and Laurasia) orogens. The so-called terrane, "Sonomia," had just collided with North America along a west-dipping subduction zone. "Stikinia"—a complex island arc now in the Cordillera from southern Alaska to southern North America—was still off the shore of the western edge of Pangaea atop a west-dipping subduction zone. A Triassic hotspot deposit, now present in the "Wrangellia" terrane (part of the Wrangell-Insular "superterrane") was forming above a rising plume. The "Alexander" terrane was above a subduction zone, inferred to be west dipping. The Siberian traps, perhaps Earth's most extensive plume-related volcanic sequence, formed in present western Siberia, between the Uralian and Altaid collisional belts. "Cimmeride terranes" (Şengör, 1984, 1987) were outboard of the Pangean continent. Bulges in the hot abyssal layer formed irregularly between deeply subducted lithospheric slabs.

In the 150–190 Ma interval (Fig. 31B), collision of the Stikine superterrane with North America occurred ca. 160 Ma, above its west-dipping subduction zone. A new subduction zone had developed, dipping beneath the "Laurasian" continent. The Atlantic Ocean had begun to open along voluminously volcanic rifted margins between Europe and North America. "Leaky-transform" style openings along the European and African continental margins later led to subsequent development of the Tethyan ophiolites.

Van der Meer et al. (2012) developed provisional maps of subduction zones in Panthalassa during the late Paleozoic and



A



B

Figure 31 (*continued on following page*). Schematic speculative cross sections of the Earth at times in the past. Symbols as in Figure 3. (A) Cross section for 250 Ma; (B) Cross section for 150–190 Ma.

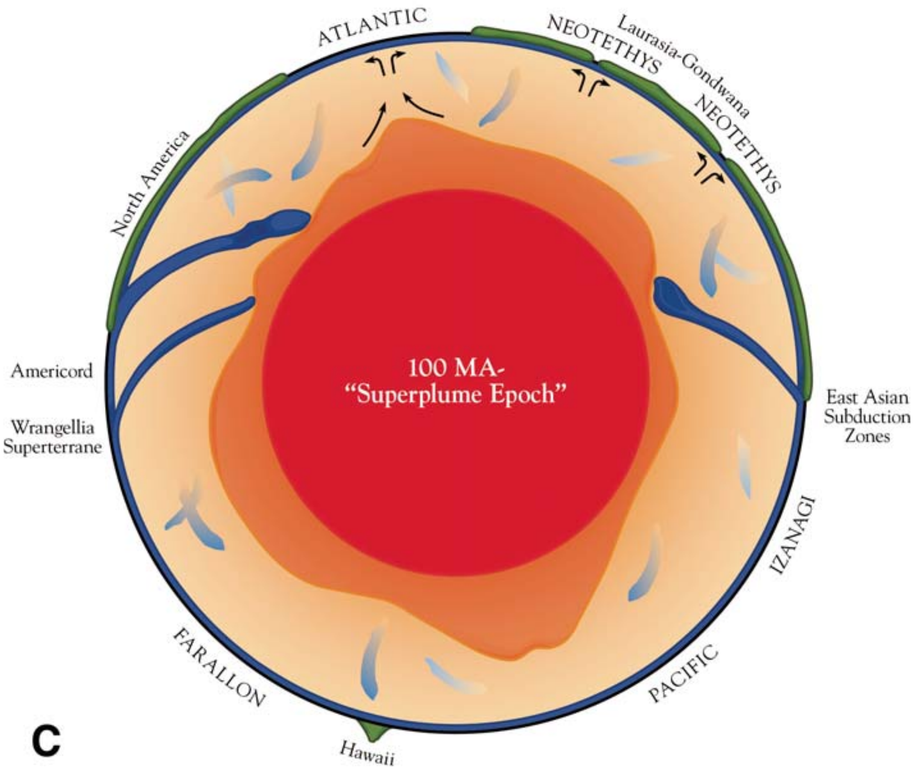


Figure 31 (*continued*). (C) Cross section for 100 Ma. See text for discussion.

early Mesozoic, using inferences from mantle tomography. Their models will help refine models developed from surface geology.

At ca. 100 Ma (Fig. 31C) the Atlantic Ocean was ~30° wide along the equator, separating North America from Europe and Africa. Narrow Neotethys rifts continued to form, and also to be emplaced along the edges of continents and microcontinents in the Alpine-Mediterranean region. Subduction zones along the Panthalassa margins of the Americas depressed the "hot abyssal layer" in those regions. Upwellings of the hot abyssal layer led to development of the Pacific plate between the Farallon and Izanagi plates.

The reconstructions imply that large upwellings may have resulted when subducted slabs pushed enriched material into piles away from the subduction zones. The production of large igneous provinces may have resulted from epochs of large-scale subduction activity.

Possible Line between Highly Magmatic Rifted Margins and Continental Assembly Episodes

The general pattern between times of major continental assembly episodes suggests a link with highly magmatic rifted margins. Approximately half of Earth's Mesozoic rifted continental margins display thick igneous sequences present as dikes, marginal plutonic complexes, and submarine seaward-dipping reflectors (e.g., Menzies et al., 2002). These highly magmatic rifted margins imply a role for plume-generated rifting. By contrast, although volcanic rocks are present along the Paleozoic rifted margins around nuclear North America (Laurentia), they are much less in volume compared with those of the Mesozoic. Abundant volcanic, possibly plume-generated, rocks were present at ca. 1100 Ma, at least in North America (e.g., Hoffman, 1988; Moores et al., 2000). This time apparently coincides with the assembly of a major supercontinent, Rodinia. Figure 32 shows a diagrammatic history of major continental convergence to form major supercontinents, their breakup, times of major large igneous provinces, thick oceanic crust, and the so-called *Mesoproterozoic Anorthosite event*. Three supercontinents are shown, the provisional *Nuna,* ca. 1.8–1.9 Ga, formed of the older cratons Slave-Tungus, and Thelon-Akitkan (of northern Canada–Siberia), Aldan (Siberian), Rae (NE Canada), Baltica (NE Europe), Superior (Canada-USA), and Wyoming (western USA) (Evans and Mitchell, 2011). These cratons are separated by 1.8–1.9 Ga orogenic belts. The implication is that they amalgamated to form the supercontinent ca. 1.8–1.9 Ga. This supercontinent may have persisted until ca. 1.4 Ga.

The breakup of Nuna led to several small continental masses that came together again ca. 1000 Ma to form Rodinia. The orogenic events representing the collision of these separate continents are generally called *Grenville-aged events* (900–1100 Ma). Rodinia persisted until its breakup ca. 750–800 Ma. Gondwana formed ca. 500–600 Ma along the Pan-African–Braziliano orogenic belts. Laurasia formed by combination of Baltica and Laurentia in the Caledonian orogeny, and then finally Pangaea formed ca. 250 Ma by amalgamation of Gondwana, Laurasia, and Siberia. The breakup of Laurasia and Gondwana led to the present world's oceans.

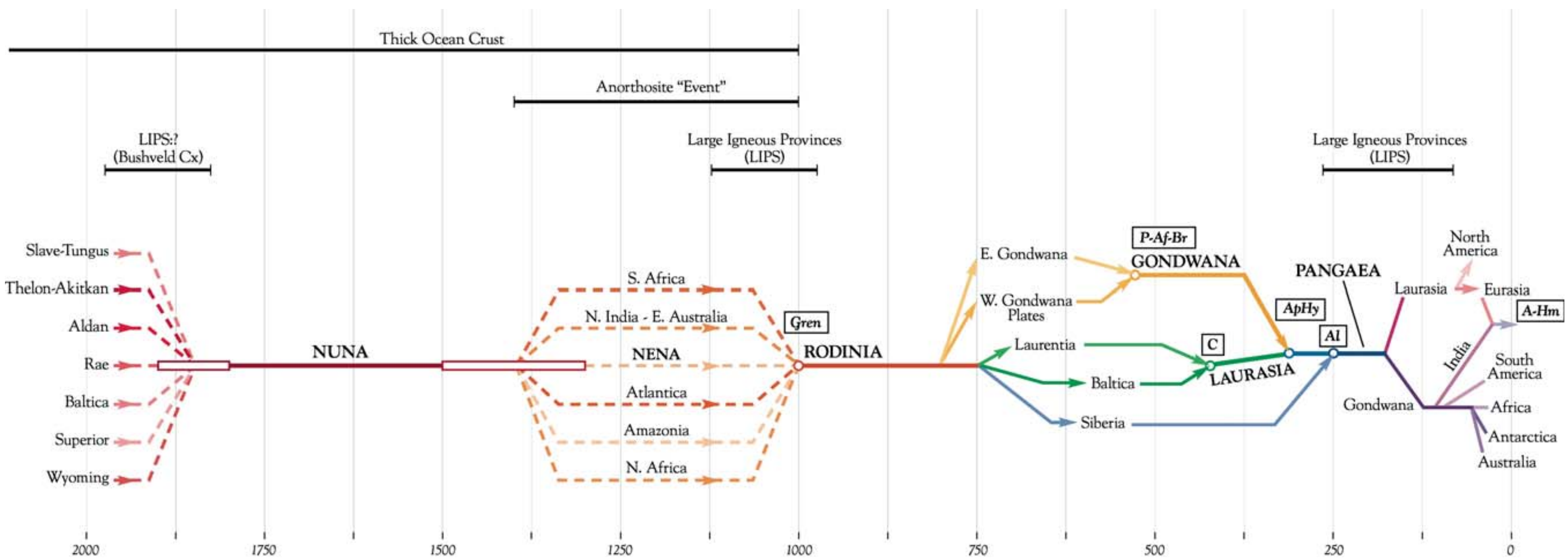


Figure 32. Schematic diagram, showing schematic evolution of continental assembly (orogenies) and fragmentation (after Moores et al., 2000). Al—Altaid-Uralian orogenies of central Asia; A-Hm—Alpine-Himalayan orogenies; ApHy—Appalachian-Hercynian orogenies; C—Caledonian orogenies; Gren—Grenville age orogenies (1000–1200 Ma); NENA—composite continent formed of Baltica, Laurentia, Eastern Siberia, and Eastern Antarctica; NUNA—proposed Paleo-Mesoproterozoic supercontinent; P-Af-Br—Pan-African–Braziliano orogenies. Diagram reads from left to right. Convergence of separate continental blocks (red) formed supercontinent Nuna, ca. 1750–1800 Ma. Nuna separated into fragments (orange) ca. 1500–1300 Ma. These fragments recombined during assembly of Rodinia ca. 1000 Ma. Rodinia fragmented ca. 775 and 750 Ma into east and west Gondwana fragments (yellow), Laurentia and Baltica (green), and Siberia (blue). Gondwana fragments recombined at ca. 500 Ma to form Gondwana. Baltica and Laurentia combined ca. 400 Ma to form Laurasia. Pangaea formed ca. 250 Ma. It fragmented to produce the present continental configuration. Heavy black lines—times of major large igneous provinces (LIPs), the "Anorthosite event" in late Mesoproterozoic time, and thick oceanic crust prior to 1000 Ma. After Moores et al. (2000).

Oceans that closed along the Alpine-Himalayan belt include the *Paleo-Tethys,* a part of the *Panthalassa* ocean that existed during the time of Pangaea.

The breakup and assembly processes outlined in Figure 32 show that the orogenic evolution of Earth involves a complex pattern. Each breakup and reassembly can be called a Wilson cycle. Just within the last 250 million years, the cycles present multiple openings and closings. The Pacific Ocean may represent a separate feature that will have a much longer history than most oceans. The pattern outlined in Figure 32 mostly concerns oceans surrounded by passive margins, currently represented by the Atlantic and Indian Oceans. Separation of Australia-Antarctica-China from Laurentia in the later Precambrian developed the Pacific Ocean, which has yet to close, although it is contracting. Will the Pacific ever close, or is it a more permanent feature of the Earth?

The diagram also shows possible plume-related activity coincident with the assembly of Nuna. For example, the Bushveld Complex, South Africa, the world's largest known layered intrusion, is dated at 2054 Ma (Scoates and Friedman, 2008). There seems to be an alternation of major convergences to form supercontinents and times of major plume activity. This relationship suggests that subducting slabs depress the hot abyssal layer in some places, causing it to rise up elsewhere in regions from which plumes emanate.

OTHER PLANETS AND MOONS

Although study of comparative planetary tectonics is still young, it has itself gone through a revolution in the past several decades. Much information is available about the tectonic features of other terrestrial planets and moons. Surface information of the various bodies improves steadily, punctuated by occasional arrival of new spacecraft, and the surface features of various planets are increasingly well known. Samples have been collected from Earth's Moon, but there is much spectral and astronomical information available on the composition of the other terrestrial bodies.

The clearest difference between Earth and Earth's Moon, Mars, and Mercury is the abundance of impact craters on the surface of the latter three bodies. Actual timing of impact events is available only from Earth's Moon (Fig. 33). Comparisons of other planets' impact density with that of the Moon allows a comparable age to be inferred for those other sites.

The Earth's Moon is thought to have been born in a collision of a proto-Earth with a Mars-sized (Stevenson, 1987; Canup and Asphaug, 2001) or smaller or larger (Halliday, 2012) object some 4.5 Ga. The Moon displays two principal surface types: lighter colored highlands and darker plains or maria, composed respectively of anorthosite and basalt. Maria mostly are ca. 3–3.5 Ga, although younger and older examples are also known (e.g., Stroud, 2009); anorthosites are older. There seem to be no surface structures of internal origin on the Moon; faults related to impact features seem to be the only identifiable tectonic features.

Mars is the most intensively studied planet after Earth, with numerous spacecraft: One, *Curiosity,* currently is exploring the Gale Crater region (see Fig. 34). Mars has a thin, CO_2-rich atmosphere and no magnetic field. Several features are of geologic interest (Fig. 34; Carr, 1984): lowlands, principally in the north polar region; highly cratered upland regions; a large rift, the 4000-km-long Valles Marineris, the longest known canyon system in the solar system (Yin, 2012); large volcanoes associated with an upland surface, the Tharsis bulge. One volcano, Olympus Mons, is the largest volcano in the solar system, 25 km high and 550 km in diameter (Carr, 1984). Complex physiographic features indicate the former presence of surface water. Observations

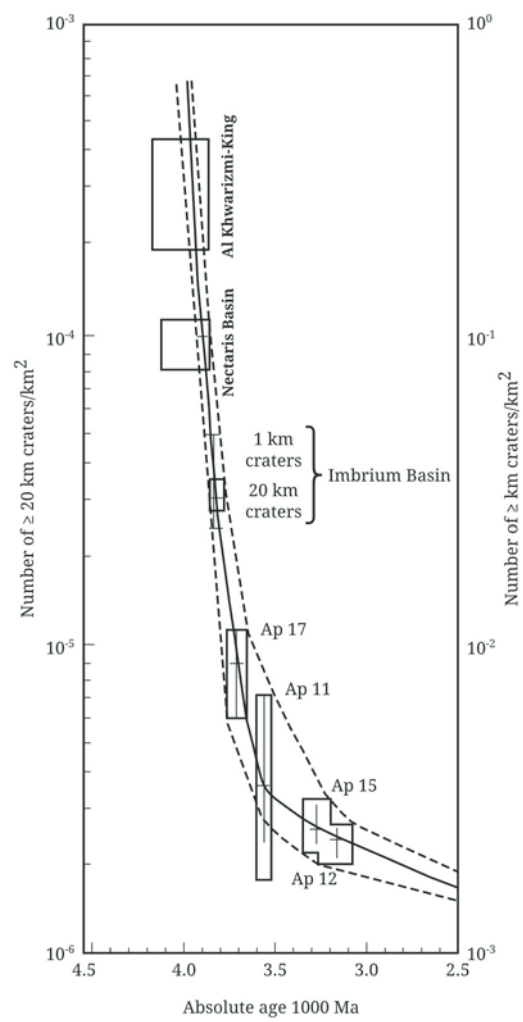


Figure 33. Estimate of number of impacts on the Moon versus time, based on samples from Apollo missions and regional crater-density surveys. Upper part of curve based upon large craters 20–40 km in diameter; lower part of curve based upon craters equal to or greater than 1 km. Connection between small and large craters in Imbrium Basin. Symbols: Ap 11 range of Apollo 11 samples and sites; Ap 21 of Apollo 12; Ap 15 of Apollo 15; Ap 17 Apollo 17. After Wilhelms (1984).

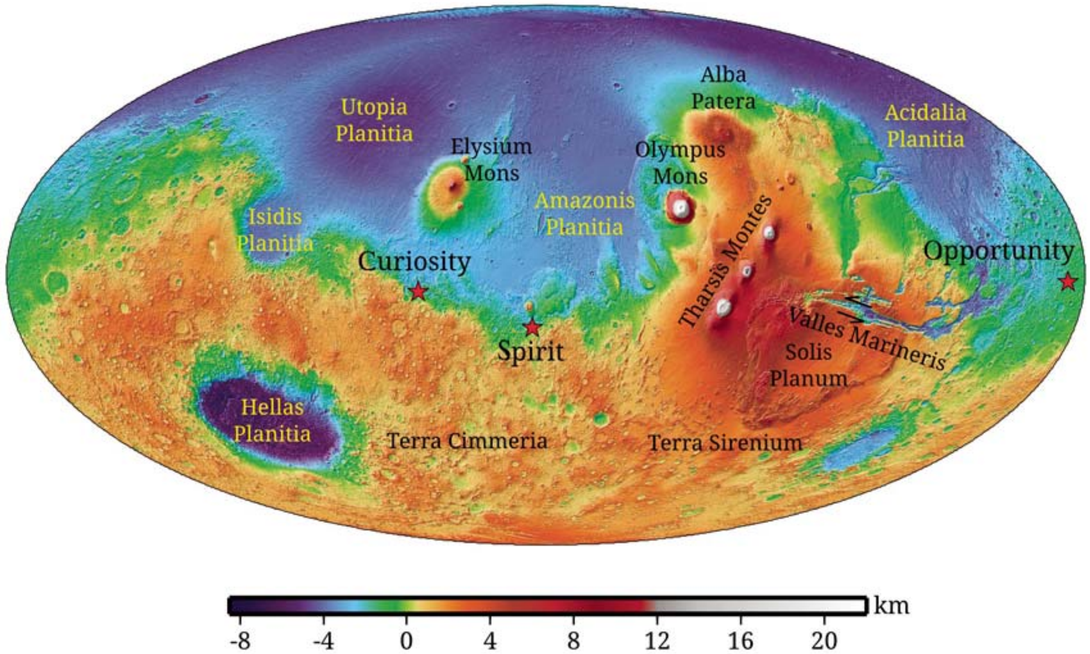


Figure 34. Relief image of Mars, showing major surface features (Smith et. al., 2003). Blue is lowland, mainly in the north polar region. Highlands are in yellows to brown. Tharsis Plateau and Olympus Mons shown in red to white. Elevation of Olympus Mons is ~25 km above median surface. Valles Marineris is ~4000 km long. Strike-slip faulting on Valles Marineris after Yin (2012).

of Mars suggest the presence of a thick, rigid lithosphere, but correlation of the Tharsis bulge with gravity and topography suggests a lack of isostatic compensation. Mars may have had an early magnetic field, but none is present now (Kerr, 2012). Yin (2012) presented evidence that the Valles Marineris is a left-lateral, strike-slip fault with ~150 km of slip.

Venus is the planet most similar in size to Earth, but conditions on the surface are quite different (McNamee et al., 1993). Venus has been subject to a runaway greenhouse condition, so a thick atmosphere shrouds the planet. The amount of CO_2 on Venus is comparable to that on Earth, but in the former it is present as CO_2 in the atmosphere, whereas on Earth most of the CO_2 is locked up in carbonate rocks. Venus tectonics consist of highlands, lowlands, intermediate ridged regions or *tesserae,* and large circular structures or *coronae* (Head and Crumpler, 1987; Fig. 35).

Venus seems to be a complex assemblage of high, multiply deformed regions (highlands) and lower volcanic plains (Fig. 35A). One highland region (Beta Regio) includes broad topographic regions with abundant shield volcanoes, extensional tectonics, and an isostatic compensation depth of ~150 km or more (Fig. 35B). This region compares in size to the East African Rift. Other features suggesting compressional structures are also abundant on Venus (Fig. 36). Complex terranes, or tesserae, may be multiply folded features. Figure 36 shows a feature in Ishtar Terra that compares in scale and apparent structural complexity with the Kashmir syntaxis, NW Himalaya and surrounding region. Crater analysis suggests that the surface of Venus is ca. 500 Ma. Large circular patterns, called *coronae,* may be either mantle upwellings or downwellings (Phillips and Hansen, 1994). However, the surface features on Venus seem to be less than one billion years old (Basilevsky and McGill, 2007).

Comparison between the elevation or hypsographic curves of Venus versus Earth shows an interesting relationship (see Fig. 37). As shown in the figure, Venus's elevation is unimodal, whereas Earth's is presently strongly bimodal. Extrapolating to the past, if Earth had a thicker oceanic crust (say, two times the present thickness, e.g., Moores, 2002, 1986), its hypsographic curve might exhibit closer spaced, subdued peaks. If Earth's oceanic crust were three times thicker than at present, then its hypsographic curve might have been unimodal. Thus Venus may represent an actualistic model for Archean-Hadean Earth. Harrison (2009) argued that the chemical nature of >4 Ga zircons from the Jack Hills, Western Australia, permit the view that Archean-Hadean Earth may have had tectonic processes not unlike those of today's Earth. If, indeed, some sort of plate tectonics were active in Archean-Hadean Earth, one might expect to see many plate boundaries and small continental (sialic) crustal masses, but a topographic form perhaps resembling that of present Venus.

Mercury, the smallest of the terrestrial planets, seems to have chiefly a highly cratered surface with evidence of contractional structures. It has a magnetic field, so a liquid core seems

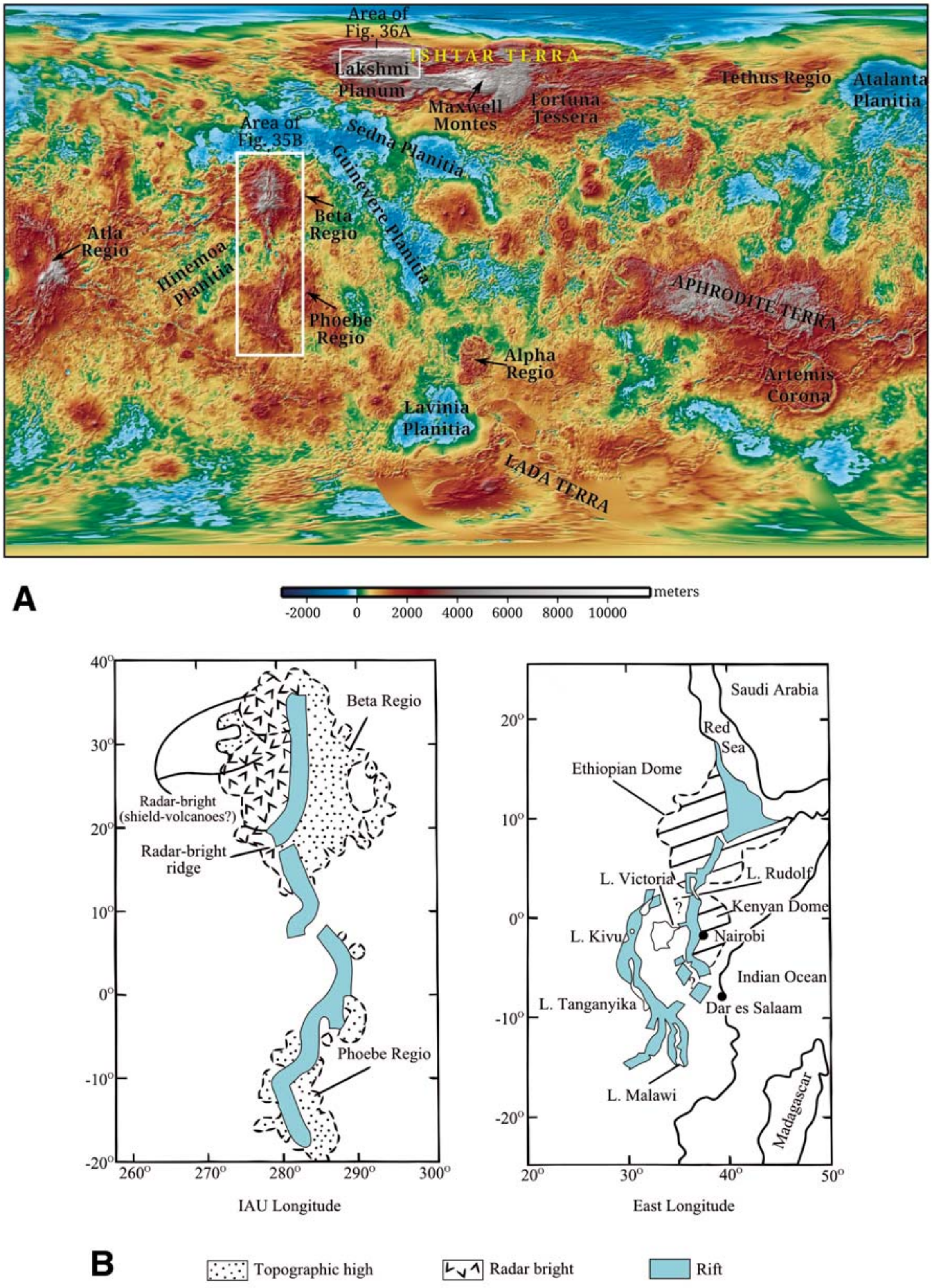


Figure 35. Images of Venus. (A) Mercator projection relief image of Venus, showing major tectonic features. Colors range from blue (lowlands) to brown and white (highlands). The Beta Regio area in western part of image is thought to resemble the East African Rift. Compressive features characterize most of the highlands. Circular structures are coronae. Redrawn after NASA Planetary Data System (ftp://pdsimage2.wr.usgs.gov/pub/pigpen/venus/Radar_properties) (B) The Beta Regio and adjacent regions of Venus, showing a possible rift valley. East African Rift region shown for comparison. After Phillips and Malin (1984).

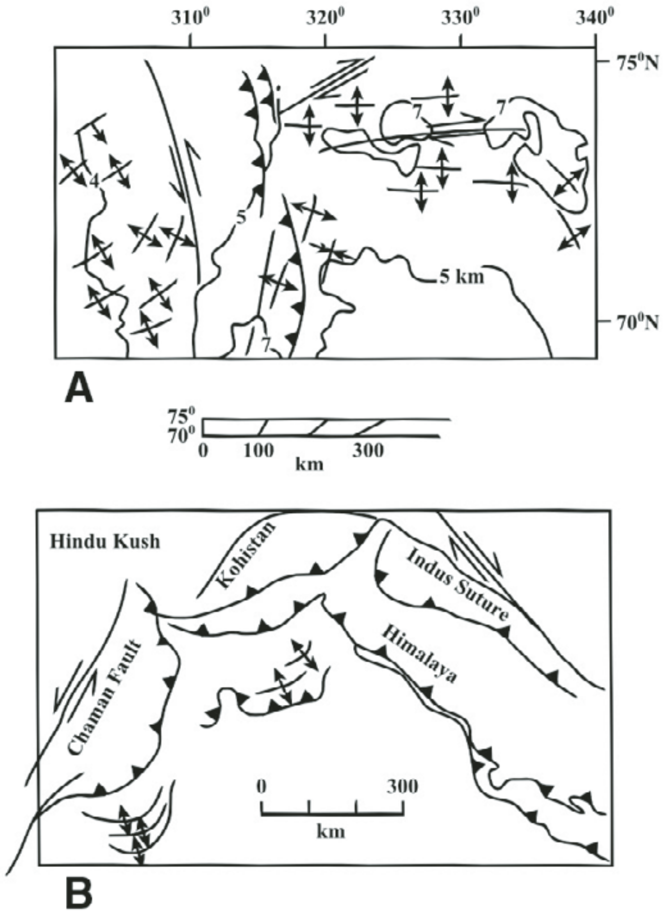


Figure 36. Possible compressive features on Venus. (A) Interpretive map of radar images with ridges of inferred compressive origin, scarps of inferred thrust-fault origin, and offset features of inferred strike-slip origin; Ishtar Terra, Maxwell Montes. (B) Comparison, at approximately the same scale, of the Kashmir syntaxis, NW Himalaya, India, Pakistan, Afghanistan. After Crumpler et al. (1986).

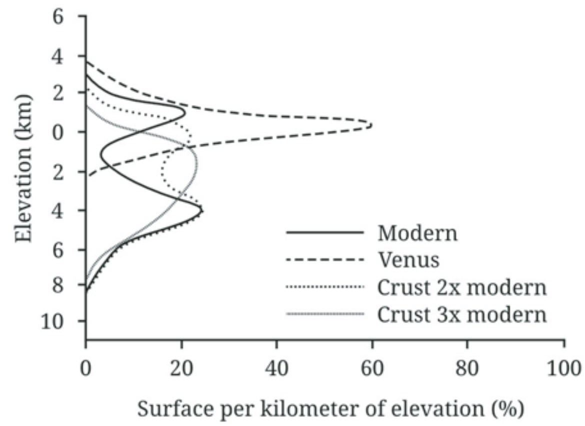


Figure 37. Comparison of modern hypsographic curves for Earth and Venus, and hypothetical curves for Earth, with oceanic crust two times modern thickness and three times modern thickness. Such crustal thicknesses may have been characteristic of Mesoproterozoic–Archean or even Hadean Earth. Redrawn after Moores (1986).

necessary (Watters et al., 2009; Kerr, 2012). Watters et al. (2012) described complex grabens and "wrinkle ridges" from Messenger spacecraft images, which they ascribe to interaction of successive lava flows and global cooling.

SUMMARY AND CONCLUSIONS

1. The Plate Tectonic Revolution of the mid–late twentieth century was a major scientific revolution that occurred simultaneously with revolutions in travel, imagery, computers, remote sensing, and presentation.
2. Earth's tectonic activity involves three types of plate boundary—divergent, conservative (transform fault), and convergent (subduction zones). All affect both continent and ocean crust. All may have occurred since 4.0 Ga. All have geologic products that may be observed along present boundaries or inferred from ancient examples.
3. Triple junctions—boundaries between three plates—are widespread; their stability is important for smooth plate evolution.
4. Collisions are interactions between thick oceanic crust or continents and subduction zones that impede movement at that boundary and cause cessation of smooth plate activity.
5. Ophiolite complexes provide on-land examples of formation of oceanic crust at divergent plate margins or in island arcs.
6. Ophiolite emplacement involves collision of a continental margin or island arc with a subduction zone, and represents a major tectonic rather than an igneous event.
7. The Earth's mantle comprises the lithosphere, asthenosphere, mesosphere, and a "hot boundary layer" above Earth's core.
8. Bulges in the hot boundary layer may give rise to plume-related igneous activity. A speculative portrayal of Earth's internal structure for the past 250 million years suggests that such bulges may arise from depressions caused by impingement of deep subducting plates.
9. Orogenic belts, the focus of tectonic attention prior to the development of plate tectonics, are scars in the Earth's crust of previous plate-tectonic activity. These belts provide insight into Earth's plate tectonic activity prior to 200 Ma.
10. There may have been at least two pre-Pangaea supercontinents—Rodinia, ca. 1000–700 Ma, and Nuna, ca. 1750–1500 Ma. These supercontinental times alternated with times of continental dispersal.
11. The Wilson cycle, the opening and closing of oceanic basins, involves sweeping together of microcontinents, oceanic island arcs, oceanic plateaus, and plume traces

into an orogenic belt as two continents converged and collided with each other.

12. The Pacific Ocean Basin apparently opened about 650 million years ago. It may be a Wilson cycle with an unusually long time frame or a different feature—perhaps part of the permanent two-thirds of the Earth's surface covered by oceanic crust.
13. Of the terrestrial planets and the Moon, only Mars shows a structure (Valles Marineris) with strike-slip displacements resembling plate tectonic movements on Earth.
14. Venus may represent a model for an Archean-Hadean Earth.

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