6.3 TECTONICS OF BASALTIC VOLCANISM ON EARTH IN THE LATTER HALF OF EARTH HISTORY

6.3.1 Tectonics of basaltic eruption at plate margins in the past

Tectonics of basaltic volcanism at past accreting plate margins: ophiolites

The idea that ophiolites represent small samples of old oceanic crust and upper mantle emplaced tectonically into orogenic belts was first clearly put forward by Dietz (1963a), Hess (1964), Gass (1968), Peyve (1969) and Laubscher (1969). An elaborate justification of this idea is not appropriate since it has been made by, among others, Thayer (1969), Coleman (1971) and Moores and Vine (1971). These and other workers have pointed out that the thickness and petrology of the internal units that compose well-developed ophiolite sequences match our present understanding of the oceanic crust very well in a general way. Although the thickness of individual ophiolite complex units may vary considerably, the range of thickness variation is comparable with layer thickness variation recorded in the ocean floors (Coleman, 1971; Dewey and Bird, 1971; Moores and Jackson, 1974).

The general sequence of rock types in ophiolite complexes has been presented above (section 6.2.1). Sediments overlying the igneous rocks are not regarded as being part of the ophiolite suite. They may consist of pelagic sediments like those covering much of the present ocean floor, or they may be clastics derived from relatively nearby continents or volcanic arcs. All but a very small proportion of the oceanic crust has been subducted. Of the crust that for various reasons has escaped this fate, most is badly shredded and dismembered; this is particularly true of the material preserved in steeply inclined orogenic zones and/or at deeper structural levels, even in Mesozoic and Tertiary orogenic belts. Thus, the general observation that recognizable ophiolite complexes become scarcer in older orogenic belts is probably explicable because more uplift and erosion have generally occurred in most parts of older belts compared with younger ones. Other reasons for relative scarcity may include the fact that earlier workers who studied older orogenic belts were unfamiliar with the utility of the ophiolite concept, the possibility that major continental collision and suturing was generally more effective and intense further back in time, and the fact that there is proportionately a smaller length of older orogenic belts exposed or preserved compared with younger ones. Because the process of ophiolite obduction is so important, a review of how it operated in a single mountain-belt is included below.

Well-preserved ophiolite complexes are not common, even in Mesozoic and younger orogenic belts. For example, the subduction of the ocean between India and Eurasia has been highly efficient, in conjunction with uplift and erosion, in removing all but a few tiny and tattered remnants of the vast ocean floor that once lay between them. Well-preserved Mesozoic and younger ophiolite complexes that have been more than cursorily described include Troodos (Wilson, 1959; Moores and Vine, 1971; Kidd and Cann, 1974; George, 1978); Vourinos (Moores, 1969; Jackson *et al.*, 1975); Canyon Mountain (Thayer, 1963; Ave'Lallement, 1976); Oman (Reinhardt, 1969; Glennie *et al.*, 1974); Antalya (Juteau *et al.*, 1977); and Papua (Davies, 1971).

Useful compendia of papers and/or reference sources on ophiolites may be found in Coleman (1977), Coleman and Irwin (1977) and the Proceedings of the International Ophiolite Symposium (Panyiotou, 1980). Well-described ophiolite complexes of Paleozoic age are known from the Appalachians and the Caledonides (Bird *et al.*, 1971; Kidd *et al.*, 1978; Dewey and Bird, 1971; Church and Stevens, 1971; Upadhyay *et al.*, 1971; Casey, 1979; Rosencrantz, 1979; Karson, 1977; Karson and Dewey, 1978; Sturt *et al.*, 1979) and less detailed descriptions of some of the widespread Paleozoic complexes in the Urals and central Asian orogenic belts are available (for example, Abdulin *et al.*, 1974; Makarychev and Shtreys, 1973).

Definite examples of pre-Paleozoic ophiolite complexes have been identified only in Pan-African orogenic belts in Morocco (LeBlanc, 1976), and Saudi Arabia (reported in Brown, 1978) and in the older Baikal orogenic belt in the U. S. S. R. (Klitin and Pavlova, 1974). Many other dismembered pieces surely remain to be identified in Proterozoic orogenic belts, since the overall geology and tectonics of both the Labrador-Cape Smith-Nelson (Wilson, 1968b) and the Coronation orogenic belts (Hoffman, 1973, 1980) so clearly involve rifting and later collision (Wilson cycles), starting in both cases about 2.15 b.y. ago. The occurrence of an extensive greenstone belt terrain of similar age to the Coronation orogen (the Birrimian of West Africa) makes it likely that ophiolites will be found eventually in greenstone belts. Extensive areas of mafic volcanics of tholeiitic compositions that are similar to compositions of oceanic crust exist in the Archean greenstone belts; what is generally lacking are equally widespread plutonic equivalents. Moores (1973) has suggested that older Precambrian oceanic crust was characterized by more abundant development of anorthosite than is younger material; this has yet to be verified. It is possible that the extensive tholeiitic submarine lavas of the greenstone belts include, besides ocean floor, and island arc-generated lavas, large thicknesses generated in the ocean by intraplate flood-basalt type magmatism.

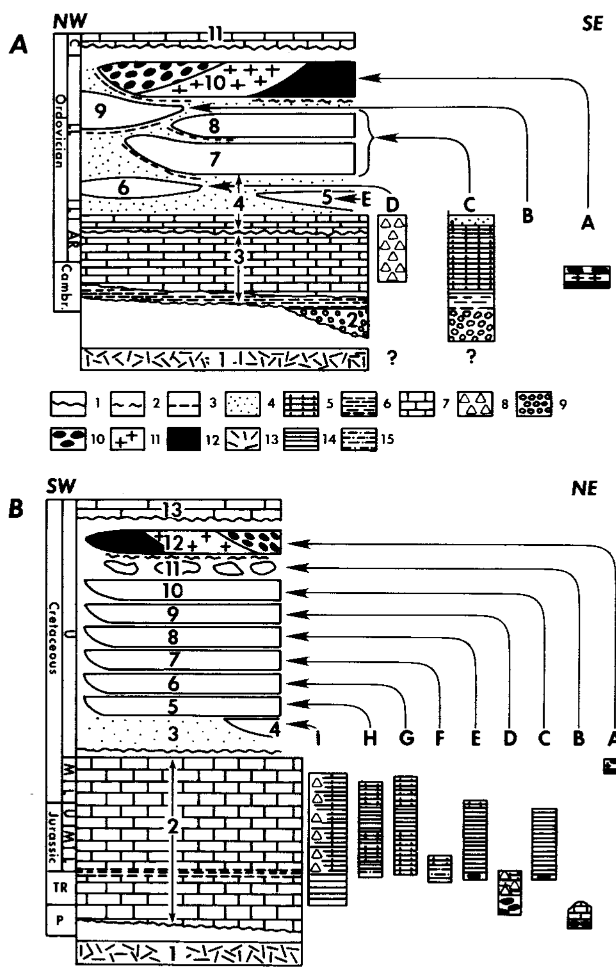
The ophiolite obduction process. In recent years there has been a great revival of interest in Alpine-type mafic/ultramafic complexes mainly as a consequence of the hypothesis that ophiolite complexes represent slices of oceanic crust and mantle generated by seafloor spreading which subsequently were emplaced tectonically into various portions of orogenic belts (Gass, 1968; Temple and Zimmerman, 1969; Moores, 1970; Moores and Vine, 1971; Church and Stevens, 1971; Dewey and Bird, 1971). This chapter takes the generally accepted view that ophiolite complexes originated as oceanic crust and mantle either at the accreting plate margins of oceanic ridges (Moores and Vine, 1971) or in marginal basins behind or between island arcs (Dewey and Bird, 1971), and were subsequently emplaced at convergent plate margins (subduction zones).

There are many problems related to the origin and emplacement of ophiolite complexes, but four are particularly difficult to resolve. First, sedimentary sequences lying upon the igneous rocks of ophiolite sequences vary greatly in character, from cherts and maganiferous lutites to volcaniclastics, and bear a variety of strikingly different relationships to the igneous rocks of the ophiolite suite, from conformable to strongly unconformable. These variable relationships are of great importance in understanding the evolution of ophiolite sequences and will be discussed below. Second, the time between igneous origin (minimum age determined by the age of sediments lying upon the pillow lavas) and tectonic emplacement may vary considerably but is often very short. This means that ophiolite generation by some form of plate accretion often is rapidly followed by either the development of a subduction zone or lateral transport to a subduction zone. Possibly related to a short time interval between ophiolite origin and emplacement are the high temperature garnet-amphibolite "aureoles" beneath the ultramafics in several transported ophiolite nappes. If portions of oceanic lithosphere were tectonically emplaced shortly after generation, they would retain much of the heat developed at their generation site and would form a thermal blanket as the ophiolite nappes moved over subjacent rocks. Several of the obduction mechanisms discussed below may provide partial solutions to this problem. The third problem is that, in many orogenic belts, ophiolites of a particularly narrow time span are preserved. For example, in the Appalachian/Caledonian System, most ophiolite complexes are of Early Ordovician age; in the western part of the Alpine System they are of Late Jurassic/Early Cretaceous age, and in the eastern part of the Alpine System in the Middle East they are of Late Cretaceous age. In both these orogenic belts it is clear from stratigraphic considerations that a much longer oceanic history was involved. Thus, although ophiolite complexex provide key insights to the oceanic origins of many orogenic belts, they do not uniquely dterminne the life spans of the ocenas involved (Dewey *et al.*, 1973). Fourth, there are great variations in the internal sequence of igneous and metamorphic rock types, although the following general sequence, from base to top, applies to all the major well-developed occurrences: dunite, harzburgite, and minor lherzolite, with strong tectonic and metamorphic recrystallization fabrics (these rocks have tectonic relationships with the country rocks); cumulate ultramafic rocks; partly cumulate, mafic, plutonic rocks; sheeted diabase; pillow lava capped by sediments. The complex variations on this basic sequence are probably a function of variation in rates of plate accretion (Dewey and Kidd, 1977) and fracture zone influence (Karson and Dewey, 1978) at oceanic ridges and in marginal basins.

Ophiolite complexes are superbly developed and exposed in the Newfoundland Appalachians. These complexes may serve as a broad but not exclusive template for other, less well-developed, Appalachian/Caledonian ophiolite sequences. The general sequence of units developed in these complexes has been presented above (section 6.2.1). There are particular problems related to the mechanisms by which ophiolite complexes are emplaced in orogenic belts and how these mechanisms may be linked to processes at convergent plate boundaries. It is implicit, in the plate tectonics model, that oceanic crust and mantle are mainly subducted in oceanic trenches. It is, therefore, of great interest to understand the mechanisms by which large slabs of oceanic crust and mantle are detached from the evolving oceanic lithosphere and tectonically incorporated into orogenic belts. Coleman (1971) has coined the term **obduction** to describe this process and to make the distinction with the usual fate (subduction) of oceanic crust and mantle.

Ophiolite complexes vary greatly in size, occur in a wide variety of tectonic positions within orogenic belts, and bear highly variable relationships to sequences against and onto which they are emplaced. They occur commonly as blocks in blueschist mélanges, olistostromes and wildflysch deposits, as highly deformed slices in steep zones of high strain, as well-preserved sequences in faulted synclines, as autochthonous or parautochthonous basement to such sequences as the Great Valley Jurassic assemblage of California, and as giant nappes transported as thin sheets over platform or exogeosynclinal sequences. The last type of occurrence has been studied in great detail in the Bay of Islands Complex, Newfoundland (Stevens, 1970; Church and Stevens, 1971; Williams, 1971), in the Semail Complex in the Oman (Reinhardt, 1969; Glennie *et al.*, 1973) and in the Papuan ophiolites (Davies, 1971). The Bay of Islands Complex and the Semail Complex bear a striking resemblance to one another in tectonic emplacement styles and tectonic/stratigraphic relationships (Figs. 6.3.1a,b). A possibly important difference between the Bay of Islands ophiolite slice and the Semail ophiolite slice is that the former faces in the direction in which the sheet was transported, so that the basal thrust contact cuts across progressively higher units of the ophiolite sequence in that direction. The reverse is true of the Semail ophiolite slice. This may reflect the initial dip direction of the thrusts along which the ophiolite slices became progressively detached from the oceanic lithosphere. Nevertheless, in both western Newfoundland and in the Oman, obducted ophiolite sheets bear very similar tectonic/stratigraphic relationships to underlying rocks and in both cases a probable continental shelf, continental rise, ocean basin paleogeography was progressively telescoped into an identical stacking order.

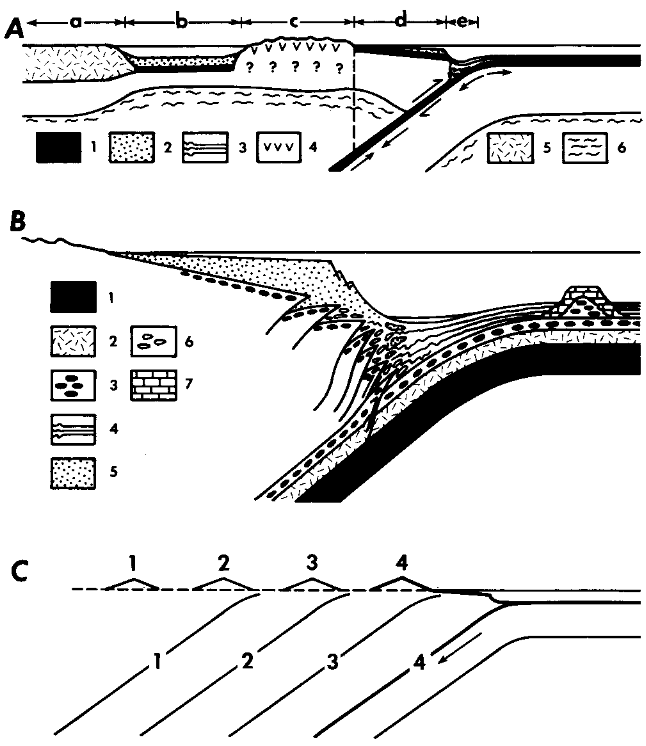
If ophiolite complexes originate as oceanic crust and mantle generated by seafloor spreading at oceanic ridges or in marginal basins, the tectonic emplacement (obduction) of ophiolite sheets and slices must involve some form of decoupling of oceanic lithosphere prior to emplacement and the expulsion of relatively dense oceanic rocks onto less dense continental rocks. The major problems are the mechanism by which this decoupling takes place, the extent to which the decoupling fractures penetrate the entire lithosphere, and the mechanism and geometry of the tectonic emplacement process (that is, the extent to which compressional vs. gravity-sliding mechanisms predominate). Several writers (Coleman, 1971; Stevens, 1970; Church and Stevens, 1971; Temple



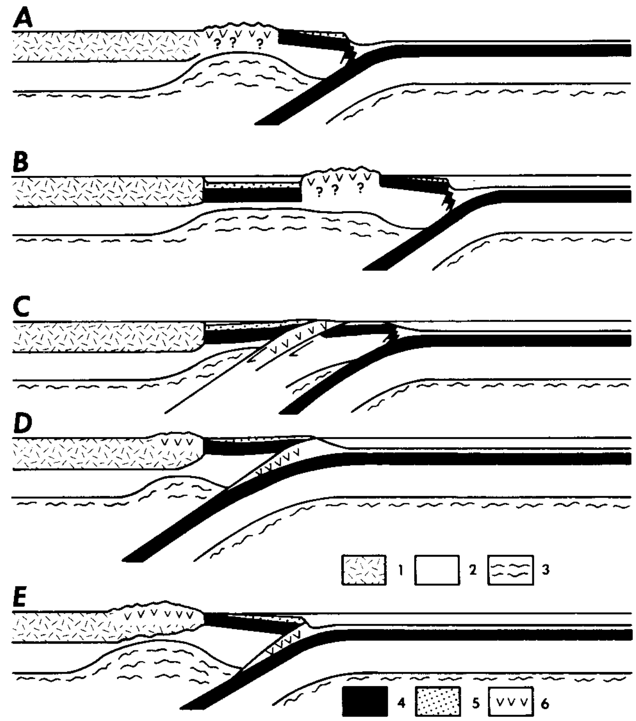
6.3.1 a. Schematic illustration of the stratigraphic/structural relationships of the Bay of Islands ophiolite sheet, western Newfoundland (data from Stevens, 1970; Williams, 1971). Key to symbols: large numerals refer to stratigraphic/tectonic units described in the text; 1—Precambrian (Grenville) crystalline basement; 2—Late Precambrian/Early Cambrian arkosic clastics and basalts; 3—Cambrian/Ordovician clastics and carbonates; 4—Ordovician carbonates and ophiolite-bearing flysch; 5—Penguin Cove allochthon; 6—Cow Head allochthon; 7 and 8—Humber Arm allochthon; 9—Coastal Complex allochthon; 10—Bay of Island allochthon; 11—Ordovician carbonates. Large letters refer to provenance and order of movement of the allochthons. Small letters in columns at left refer to stratigraphic age; AR—Arenigian; L—Llanvirnian; LL—Llandeilian; C—Caradocian. Small numerals in legend: 1—unconformity; 2—garnet-amphibolite; 3—ophiolite-bearing mélange; 4—ophiolite bearing flysch; 5—calcareous flysch; 6—shallow marine clastics; 7—shallow marine carbonate; 8—carbonate conglomerate and breccia; 9—red and green arkose, greywacke, shale and basalt; 10—pillow lava; 11—gabbro; 12—ultramafic rocks; 13—crystalline silicic basement rocks; 14—radiolarian chert; 15—deep-water quartz-rich clastics. **b**. Schematic illustration of the stratigraphic/structural relationships of the Semail ophiolite sheet, Oman. Data from Glennie *et al.* (1973). Key to symbols: large numerals refer to stratigraphic/tectonic units described in the text; 1—Precambrian crystalline basement; 2—Permian to Cretaceous carbonates; 3—Cretacrous flysch; 4, 5, 6, 7, 8, 9, 10 and 11—Hawasina allochthons; 12—Semail allochthon; 13—Upper Cretaceous carbonates. Large letters refer to provenance and movement order of the allochthons. Small letters in columns at left refer to stratigraphic age; P—Permian; TR—Triassic.

and Zimmerman, 1969; Dewey and Bird, 1970, 1971; Williams, 1971) have discussed these obduction problems and have offered various kinds of solutions. These solutions, among others, are discussed below.

Upwedging in subduction zones. Upwedging is a process that may occur beneath and behind the inner walls of oceanic trenches (subduction zones), whereby slices of oceanic crust and mantle are ripped from the upper part of the descending plate and then wedged and packed in high-pressure mélanges (glaucophane, lawsonite, jadeite assemblages) against the leading edge of the other plate (Fig. 6.3.2). Seismic first motion studies



6.3.2 **a**. Schematic section from a continental margin (a), across a rear-arc oceanic basin (b), volcanic arc (c), arc-trench gap (d), trench (subduction zone, e) to an ocean. 1—oceanic crust; 2—volcanogenic sediments; 3—oceanic sediments; 4—silicic/intermediate volcanics; 5—continental crust; 6—low-velocity zone. **b**. Schematic illustration of some possible relationships at a subduction zone. 1—ultramafic rocks (mantle); 2—gabbro (oceanic layer 3); 3—pillow lavas and diabase (oceanic layer 2); 4—argillite and chert (oceanic layer 1); 5—volcanogenic sediments of the arc-trench gap; 6—blueschist and ophiolite mélanges; 7—carbonate fringing a submerged seamount. **c**. Diagrammatic illustration of continental accretion by the successive oceanward movement of the site of subduction. 1, 2, 3—former positions of volcanic arc and Benioff Zones; 4—present position of volcanic arc and Benioff Zone.



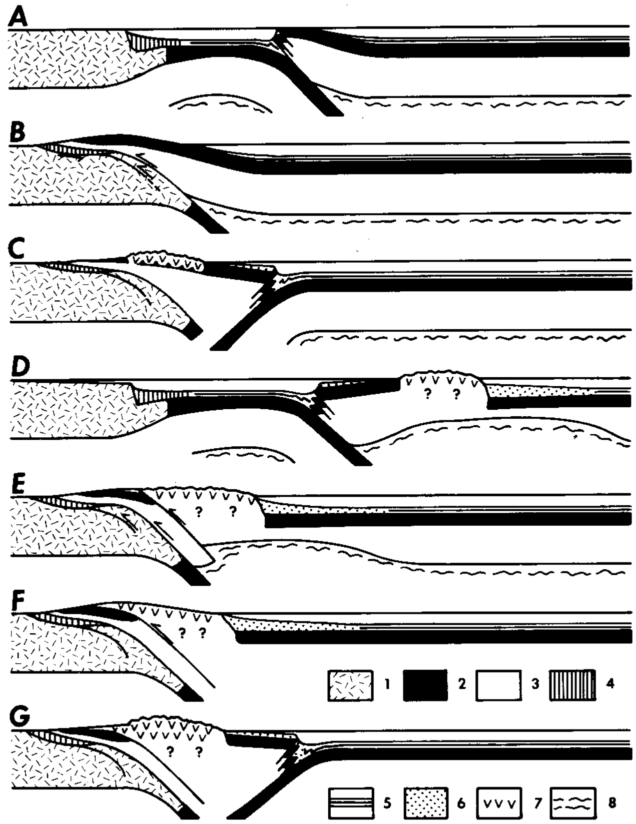
6.3.3 Schematic illustration of Blake and Jones (1973) hypothesis of the obduction of rear-arc basin ophiolites by the subduction of the associated volcanic arc. 1—continental crust; 2—mantle; 3—low-velocity zone; 4—oceanic crust; 5—volcanogenic sediments; 6—silicic/intermediate volcanics.

(Isacks and Molnar, 1969) have shown that the upper part of the descending plate is in tension where it bends to descend, and that thrust fault solutions characterize the region below the inner walls of trenches (Fig. 6.3.2). Weakening and cracking of oceanic crust and upper mantle is likely to occur in the tensional regime, following which it becomes subjected to strong shearing forces in which slabs of ophiolite could become detached and incorporated into mélange sequences. The Mesozoic Franciscan mélange assemblage contains ophiolite slabs of varying sizes believed by some authors (Hamilton, 1969; Dewey and Bird, 1970) to have been obducted in this way. At low structural levels, such ophiolite slices are likely to be encased in tectonic blueschist mélanges; at high levels they may be tectonically overlain by olistostrome and flysch sequences (Fig. 6.3.2).

It is now supposed that many arc-trench gap terranes (Dickinson, 1971) have an ophiolitic basement (Blake and Jones, 1973; Dewey and Bird, 1971). This is clearly the case for at least the outer margin of the Great Valley of California. During the Cretaceous, this terrane lay between the Franciscan subduction zone and the Sierran arc. Along the outer margin of the Great Valley, Late Jurassic ophiolites and their cover of Jurassic sediments were thrust westwards across various flysch and mélange terranes of the Franciscan (Bailey *et al.*, 1970). Blake and Jones (1973) have argued for a new and important mechanism to explain the relationship between the Great Valley and the Franciscan (Fig. 6.3.3). From a consideration of the Late Jurassic age of the Great Valley ophiolite basement, and the stratigraphy and geometry of successive Franciscan belts, they have argued that the Great Valley originated by rear-arc basin spreading and that, during the Cretaceous, the arc was largely subducted beneath the Great Valley during the inception of the Franciscan subduction zone (Fig. 6.3.3). Fragments of this supposed Late Jurassic volcanic arc are preserved in the Franciscan. This hypothesis explains the short time-span between origin and emplacement of the Great Valley ophiolite basement.

Progressive packing of ophiolite slices and arc fragments against the leading edge of a continent may continue over a long period of time and lead to a form of continental accretion (Fig. 6.3.2), perhaps exemplified by the Klamath Mountains of northern California and southern Oregon. Here, a series of ophiolitic thrust plates are stacked against one another and the plates become successively younger westwards. They are intruded by arc-type igneous rocks, which are probably the result of the progressive or intermittent westward-migration of volcanic arcs as the subduction zones migrated westwards, so that new volcanic arcs developed on older obducted ophiolite foundations.

Obduction by compressional telescoping onto Atlantic-type continental margins. The simplest form of this type of obduction may follow from the development of a subduction zone near the continental margin with the geometry shown in Fig. 6.3.4. Above and behind the subduction zone, a welt of oceanic crust and mantle rides up over the descending plate. The ocean, intervening between the continental margin and the subduction zone, is progressively swallowed until the continental margin arrives at the subduction zone and a giant wedge or slice (nappe) of oceanic crust and mantle is pushed across the continental margin (Fig. 6.3.4). McKenzie (1969) has argued that buoyancy of the relatively light continental crust is likely to prohibit its extensive subduction; therefore, a flip in subduction polarity will occur so that the continental margin with its obducted ophiolite sheet will now lie above a descending plate (Fig. 6.3.2). This mechanism should yield a distinctive arrangement of tectonic and stratigraphic units. The stacking order should be similar to that observed in



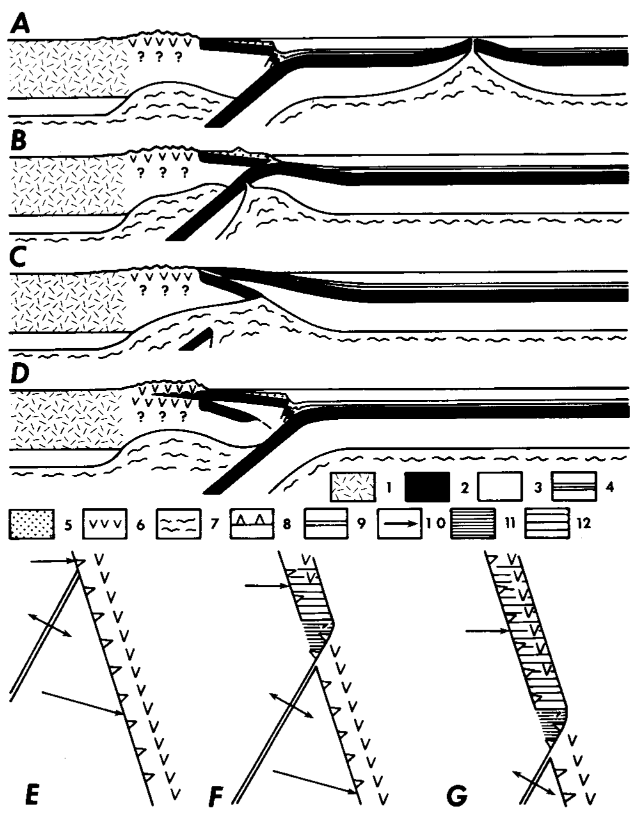
6.3.4 a,b,c. Obduction of an ophiolite sheet by the partial subduction of an Atlantic-type continental margin. **d,e,f,g.** Obduction of an arc-trench gap ophiolite sheet by the collision of an Atlantic-type continental margin with a volcanic arc. 1—continental crust; 2—oceanic crust; 3—mantle; 4—continental rise sediments; 5—oceanic sediments; 6—volcanogenic sediments; 7—silicic/intermediate volcanics; 8—low-velocity zone.

western Newfoundland, with a giant ophiolite sheet overlying oceanic, continental rise, and continental margin facies allochthons. The movement and assembly order should be the reverse of the stacking order. This model also yields the corollary that blueschist, ophiolite-bearing, subduction-mélanges should intervene, at least in the initial stages of nappe assembly, between the high-level ophiolite slice and subjacent continental rise slices. As movement progresses, these blueschist mélanges could be progressively attenuated into thin metamorphic "aureoles" beneath the ophiolite nappe and should indicate a protracted polyphase history of subduction and obduction tectonics. Complex relationships are likely to occur between different kinds of flysch, olistostrome, wildflysch, and tectonic mélanges. Early subduction mechanisms, involving low-level tectonic mélanges and high-level ophiolite-bearing olistostromes in the trench, may progressively feed flysch and olistostrome wedges advancing up the continental rise. These will subsequently be transported as new allochthonous sheets are successively detached from their basement in advance of the moving ophiolite sheet. Finally, the allochthon moves into foreland flysch basins on top of the former continental shelf.

The extent to which the high-level ophiolite sheet carries a cover of pre-obduction arc volcanics will depend upon the history of pre-obduction subduction. If only a short time intervenes between initiation of subduction and final obduction, arc volcanics may be scarce or absent—for example, Papua (Davies, 1971). If, however, a large tract of ocean intervenes between the continental margin and the subduction zone, a fully developed trench, arc-trench gap, arc, and rear-arc basin (Fig. 6.3.4) may eventually arrive at and collide with the continental margin. In this case, an arc-trench gap ophiolite basement, with a cover of arc-trench gap sediments, will be obducted onto the continental margin with a subjacent sequence of blueschist mélanges and continental rise allochthons (Fig. 6.3.4). Further convergence may lead to overthrusting of the volcanic arc assemblage, as a high-level nappe (Fig. 6.3.4). This may be followed by flipping of subduction polarity so that the obducted sequence is now underthrust from the oceanward side (Fig. 6.3.4). This may yield a situation where a new assemblage of arc volcanics and intrusives develops on and invades the obducted complex, and a new arc-trench gap, with a young rear-arc ophiolite basement, develops on the oceanward side.

This general mechanism, although providing an effective way of obducting ophiolite sheets, can yield an exceedingly complex series of relationships between continental rise allochthons, flysch and mélange sequences, volcanic arcs and ophiolites obducted by different mechanisms and originating in different ways. In particular, this mechanism does not yield an *a priori* time relationship between generation of oceanic crust and mantle by seafloor spreading and its subsequent obduction. If an arc were partially subducted beneath the rear-arc basin just prior to, or during, obduction onto the continental margin, the young obducted ophiolite sheet with its rear-arc basin sediment cover (andesitic flysch?) would be separated from the continental rise allochthons by a smeared-out volcanic arc. Such a smeared-out arc could have an older ophiolite basement which has been complexly deformed, metamorphosed, and injected by arc intrusives.

Obduction by gravity sliding onto Atlantic-type continental margins. This is a concept advocated by Reinhardt (1969) for the emplacement of the Semail Complex and argued by Church (1972) and Church and Stevens (1971) for the emplacement of the Bay of Islands sheet in western Newfoundland. It involves the progressive uplift of an actively spreading oceanic ridge, the detachment of slices from the *upper part* of the lithosphere and the subsequent gravity sliding of these slices onto the continental margin (Fig. 6.3.5). This mechanism could explain basal contact aureoles, since the young upper lithosphere would still be hot during emplacement. It would also account for a short time interval between the origin by seafloor spreading and tectonic emplacement. It could further explain why, in the Semail Nappe, the ophiolite sequence faces in the opposite direction (Fig. 6.3.1b) from the nappe transport direction. It seems to the writers, however, that this mechanism does not explain the progressive detachment of underlying nappes. It is also difficult to accept gravity



6.3.5 a,b,c. Obduction of an ophiolite sheet onto an Atlantic-type continental margin by gravity sliding from an uplifted oceanic ridge. **d,e,f,g.** Obduction of an ophiolite sheet onto an Atlantic-type continental margin by the transformation of a spreading oceanic ridge to a subduction zone. 1—continental crust; 2—oceanic crust; 3—mantle; 4—continental rise sediments; 5—oceanic sediments; 6—volcanogenic sediments; 7—silicic/intermediate volcanics; 8—low-velocity zone; 9—garnet-amphibolite.

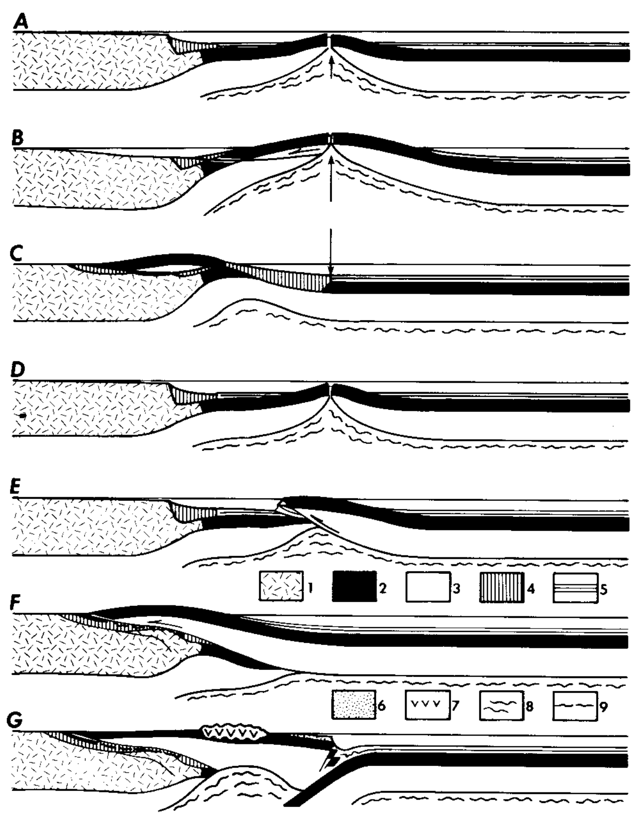
sliding for a distance of 1200 km as claimed by Glennie *et al.* (1973). Last, it is difficult to visualize or account for the pre-gravity-sliding topographic inversion.

Obduction by transformation of a spreading ridge to a subduction zone adjacent to an Atlantic-type continental margin. This model is illustrated in Fig. 6.3.5. The change from a spreading plate boundary to a subduction plate boundary may result from rapid rearrangements of relative plate motion. However, it is more likely that it results from a progressive change in position of the plate boundary (with respect to the instantaneous rotation pole describing the relative motion of the two plates), mandated by the evolution of rigid plates on a sphere (Dewey, 1975). Whatever the causes of the relative change in motion, slices of young, hot, oceanic crust and mantle may be obducted onto the continental shelf. In many respects, a tectonic/stratigraphic arrangement of stacked allochthons similar to that developed in the second model would result, except that subduction mélanges are likely to be absent, and thermal aureoles could be extensively developed beneath the obducted ophiolite sheet. This model would yield a short time span between origin and tectonic emplacement of the obducted ophiolites.

A similar possibility could result from the conversion of a transform boundary, with short ridge segments parallel to the continental margin, to a subduction zone. Such a relationship, which may have occurred during Late Cretaceous times, would explain the origin and emplacement of the Semail ophiolites (Dewey *et al.*, 1973). The oceanic crust and mantle generated along the short ridge segments may be preferentially obducted, due to their higher topography and heat flow. This would yield localized obducted ophiolite sheets along the continental margin.

Obduction by the interference of a spreading ridge and a subduction zone. Consider the situation illustrated in Fig. 6.3.6a–d, where a spreading ridge approaches a subduction zone. Eventually, the ridge impinges on the subduction zone, at which time there will develop a complex interaction of subduction-related tectonic-sedimentary, and spreading-related tectonic-igneous activity. At this stage, one of several things may happen. Following the swallowing of the intervening plate, the decaying ridge may be subducted (Fig. 6.3.6). Alternatively, it is conceivable that a hot ridge flank rides upward, across the trench, to be obducted onto the arc-trench gap and arc terranes as a hot ophiolite slice. This would yield a very short interval between origin and emplacement of the ophiolite sheet, but this model cannot be used to explain obduction onto Atlantic-type continental margins. It seems unlikely that the ridge and trench intersections with the continental margin would exactly parallel the margin. If the intersections were oblique, ophiolite sheets with successively younger sea-floor spreading ages could be progressively emplaced across the arc diachronously (Fig. 6.3.6e–g).

Obduction from rear-arc oceanic basins. It has been suggested (Dewey and Bird, 1971) that a common form of ophiolite obduction is related to the closure of rear-arc marginal basins, and that during such closure, by subduction, slices of oceanic crust and mantle may be expelled onto adjacent continental forelands and emplaced as ophiolite sheets. Packham and Falvey (1971) have suggested that, following the generation of a rear-arc basin by the migration of a volcanic arc away

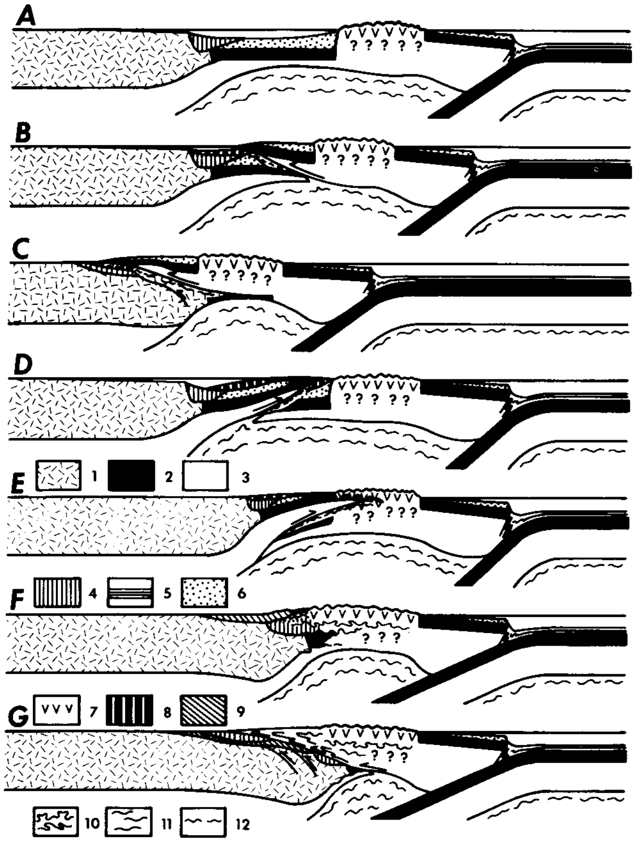


6.3.6 Various possible geometric relationships resulting from the progressive interaction of a subduction zone and a spreading ridge. 1—continental crust; 2—oceanic crust; 3—mantle; 4—oceanic sediments; 5—volcanogenic sediments; 6—silicic/intermediate volcanics; 7—low—velocity zone; 8—subduction zone (triangles on leading plate edge); 9—oceanic ridge; 10—direction of motion of a plate with respect to an adjacent plate (length of arrow proportional to rate of relative motion); 11—ophiolite sheet undergoing obduction; 12—obducted ophiolite sheet.

from its adjacent continental margin (Karig, 1971a), it becomes progressively filled with thick clastic sequences derived from both the continent and the arc. As the sediment pile thickens, anatectic fusion is supposed to occur as a result of thermal blanketing. The marginal basin is supposed to collapse and telescope with remobilized mantle peridotites flowing up as mobile screens and extrusion nappes moving over the continental margin. This model, therefore, involves the shortening, deformation and metamorphism of an oceanic tract without subduction. It is difficult to visualize how, in this model, intact ophiolite nappes become obducted, although the model could explain many serpentinite screens in high grade metamorphic terranes. Dewey and Bird (1971) have suggested that ophiolite sheets may originate from a rising welt in advance of a subduction zone, similar to that of the Mediterranean Ridge (see discussion below).

Another mechanism could account for the obduction of thin, hot sheets of oceanic lithosphere from marginal basins, which would explain most of the more problematical features of the emplacement of, for example, the Bay of Islands ophiolite complex. In the high heat flow region of a volcanic arc and rear-arc basin, the lithosphere is particularly thin (Fig. 6.3.2). The thin oceanic marginal basin lithosphere may preferentially fail along gently dipping thrust surfaces if a compressional stress field is applied to the region. Under these circumstances a thin sheet of marginal basin lithosphere may become detached and begin to ride over adjacent lithosphere, finally to become emplaced as a thin ophiolite sheet on the adjacent continental foreland (Fig. 6.3.7). This is basically a form of plate convergence whereby a thin, initially hot, but progressively cooling sheet of oceanic lithosphere is obducted over progressively cooler and thicker lithosphere.

Initially, the sheet would be attached to its hot mobile infrastructure (low-velocity zone), but the zone of dislocation would gradually become localized along a progressively cooler and thinner zone of high shearing strain. During the early stages of detachment, primitive mantle materials of the low-velocity zone (lherzolite and ariégite?) could be dragged along, and could partially lubricate the movement zone. As they cooled, these rocks would move into cataclastic fields of strain and could be represented by the lherzolites and ariégites at the base of the Bay of Islands ophiolite slice. During the initial phases of movement, the hot sheet could peel off and dynamothermally metamorphose thin slices of the mafic oceanic crust and its sediment cover. As the cooling sheet moves out across the inner margin of the basin onto the continental foreland, exotic metamorphosed (garnet amphibolite?) mafic slivers would be carried along at the base of the sheet and become sandwiched between the ophiolite sheet and continental-rise nappes successively detached from their roots. Because heat for the metamorphic reactions is derived from the overlying hot sheet, inverted prograde metamorphic gradients would characterize these transported mafic slices. During the early prograde metamorphic phase, fluids driven from wet sediments below the moving sheet would lubricate the shearing zone and reduce frictional resistance to movement. Progressive movement and cooling of the ophiolite sheet would lead to progressively semibrittle then to brittle conditions so that, during the late and final stages of emplacement, retrograde metamorphism and cataclastic textures would become superimposed upon earlier prograde ductile textures. Serpentinite



6.3.7 **a, b, c**. Obduction of an ophiolite sheet onto a continental margin from a rear-arc basin. **d, e**. Obduction of an ophiolite sheet onto a volcanic arc from a rear-arc basin. **f, g**. Obduction of a silicic crystalline sheet. 1—continental crust; 2—oceanic crust; 3—mantle; 4—continental rise sediments; 5—oceanic sediments; 6—volcanogenic sediments; 7—silicic/intermediate volcanics; 8—ophiolitic flysch; 9—exogeosynclinal flysch and molasse; 10—polyphase deformation and metamorphism; 11—low-velocity zone; 12—garnet amphibolite.

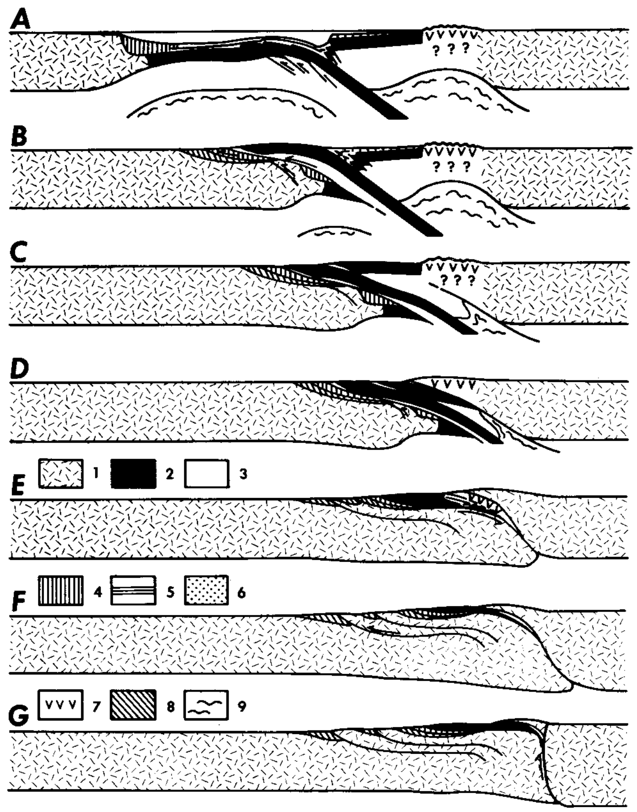
mylonites would be formed along the base of the ophiolite sheet and rodingite mineral assemblages would be developed along the contact zone in the underlying metamorphic mafic slivers.

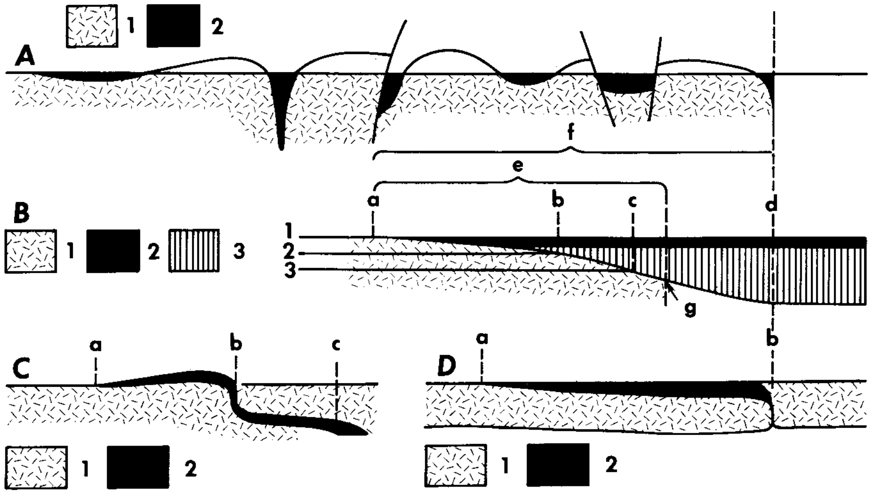
There could be a whole range of transitions from rear-arc basin ophiolite obduction to the backthrusting of continental rocks over continental lithosphere depending upon the width, or absence, of rear-arc basins. Where rear-arc basins are absent, sheets of silicic crystalline rock could be thrust back across the foreland (Fig. 6.3.7). In addition, there is no *a priori* reason why ophiolite sheets should not be driven from the marginal basin onto the associated volcanic arc (Fig. 6.3.7). This model has the particular advantage of accounting for a short time interval between the generation of oceanic crust and mantle by spreading and its emplacement by obduction.

Obduction by continental collision. The various ocean basins of the Mediterranean region are believed by some authors (Smith, 1971; Dewey *et al.*, 1973) to be remnants of oceans generated at different times by seafloor spreading. The Alpine Orogenic Belt, in addition to these remnant oceans, is believed to register a complex history of plate interactions during the general convergence of the Eurasian and African Plates. This process of convergence is continuing at the present day. Several features of the Mediterranean may provide clues to the processes leading to ophiolite obduction during the final stages of ocean closure that eventually lead to continent/continent collision.

Rabinowitz and Ryan (1970) have argued, from gravity and seismic-profiling data, that the Mediterranean Ridge has a core of upthrust basement wedges of oceanic crust and mantle, which are wedging into and deforming the overlying oceanic sediment pile. This ridge lies between the Ionian-Cretan trench, where subduction of the Levantine Ocean is occurring, and the advancing stable north African continental margin. Dewey and Bird (1971) have suggested that continued convergence in this tectonic situation could lead to the oceanic basement wedges being driven (obducted) onto the continental margin. A possible sequence of events illustrating this obduction model is shown in Fig. 6.3.8. As the rising wedges of oceanic crust and mantle rise and are caught in the jaws of the continent/continent vise, they detach and begin to move up the advancing continental rise, detaching and transporting underlying allochthonous sheets (Fig. 6.3.8). Such an ophiolite sheet is likely to be detached, transported and emplaced as a relatively cold nappe so that there is less potential for such a sheet to develop subjacent thermal metamorphism. Continued convergence may lead to the overthrusting of the arc-trench gap (perhaps as a higher ophiolite sheet) and, eventually, to overthrusting of the metamorphic plutonic and volcanic rocks of the volcanic arc (Fig. 6.3.8). This sequence would generate the following stacking order from base to top: continental rise allochthons lying on exogeosynclinal flysch and wildflysch mélanges; a lower ophiolite sheet; attenuated blueschist mélanges; a higher ophiolite sheet; and a high-level crystalline sheet of arc metamorphics, plutonics, and volcanics.

Following total subduction of oceanic tracts, and perhaps a minor amount of continental subduction, continuing convergence may lead to a further sequence of intracontinental mechanisms of crustal shortening. Initially, crystalline basement nappes may break loose in a progressive sequence away from the main suture zone (Figs. 6.3.8–9). During the early parts of this phase the ophiolites are expelled, in a downward-narrowing





6.3.9 **a**. Preservation of ophiolite occurrences, which once formed parts of a continuous obducted sheet, by post-obduction folding and faulting. 1—continental crust; 2—ophiolites. **b**. Schematic illustration of the rooting of a continuous obducted ophiolite sheet. 1—continental crust; 2—pillow lava, diabase and gabbro; 3—ultramafic rocks. **c**. Large scale monocline in obducted ophiolite sheet. 1—continental crust; 2—ophiolites. **d**. Obducted ophiolite sheet. 1—continental crust; 2—ophiolites.

envelope, from the main suture zone. At deep levels, the suture may be steeply dipping, cryptic and reduced to a narrow zone of high strain (Dewey and Burke, 1973). At high levels, away from the suture, the ophiolites may be the highest unit of the stacking sequence but will be overlain by a crystalline nappe unit where the suture zone flattens above its cryptic root. Continued continental convergence beyond this phase may lead to a stage where the whole continental crust begins to thicken so that potassic/silicic partial melts originate in the lower crust (Dewey and Burke, 1973). These potassic/silicic liquids rise to form late-stage potash-granite intrusions and dacitic/rhyolitic ignimbrite calderas. Finally, isostatic readjustment may lead to high-angle back-thrusting (*retrocharriage*) of the nappe units (Fig. 6.3.9).

No single model is likely to account for the emplacement of all obducted ophiolite sheets. Ophiolite obduction is probably a process involving several—if not many—mechanisms. These may include some of the models discussed above. Which, if any, of these models was responsible for the emplacement of particular ophiolite occurrences can only be answered by detailed field studies of their structural and stratigraphic relationships. In any particular orogenic belt, great complexities may arise from different basic obduction templates being assembled in a variety of temporal sequences. Great geometric complexities may arise as a result of irregularities in the colliding margins of continents, and the diachronous evolution of plate boundaries and triple junctions (Dewey, 1975). Projections in approaching continental margins will collide first and result in intense suturing; salients may never collide and may preserve oceanic remnants buried beneath thick sedimentary sequences (Dewey and Burke, 1973). If orogenic belts result from "cycles" of ocean opening and closing, large parts of the geologic record will be destroyed. This includes not only the obvious subduction of oceanic lithosphere but also more subtle possibilities such as the subduction of volcanic arcs, as suggested by Blake and Jones (1973).

We are, therefore, faced with the problem of reconstructing complex events from an incomplete record. Many of these lost relationships may be critical to a full understanding of orogenic evolution. Although plate tectonics has in many ways increased our understanding of orogenic evolution, and thereby invigorated studies of orogenic belts, it has given us fresh insights into the kinds of questions to which we may never know the answers, and made us aware of the bewildering array of complexities inherent in orogenic evolution. For example, there has been discussion (Cady, 1972) on whether the Appalachian/Caledonian orogenic belt evolved dominantly on a sialic or an oceanic basement. The present distribution of sialic basement rocks, according to the plate tectonics model, is a poor guide to the nature of the basement during the total evolution of the orogen, because ocean closing must involve subduction and, therefore, the destruction of oceanic basement and the preservation of continental basement. Furthermore, the age of origin and emplacement of ophiolite complexes may be the worst possible guide to the total oceanic history involved in the evolution of an orogenic belt. It cannot be supposed that the age of the oldest ophiolite complex in an orogenic belt designates the oldest possible age of oceanic tracts involved in the history of that belt, and that the age of ophiolite obduction always marks the total destruction of those oceanic tracts. The possibility of rear-arc and interarc oceanic basins must be considered. Ophiolites generated in such basins will be considerably younger than the oldest oceans involved in the evolution of the orogen, and the time between origin and emplacement may be very short. The oldest oceanic tracts may be completely destroyed and their existence may be argued only from other facies (Bird and Dewey, 1970; Dewey and Bird, 1971) and geometric considerations (Smith, 1971; Dewey *et al.*, 1973).

The writers suggest that rear-arc basin spreading, with the volcanic arc moving away from the continental margin, followed shortly thereafter by the obduction of marginal basin ophiolites, appears to offer solutions to the following three particular ophiolite problems. First, as mentioned above, ophiolite origin and emplacement are commonly separated by only a short time span. Second, as Sclater *et al.* (1971) have shown, there is a clear relationship between age (and heat flow) and elevation of the ocean floor. Obduction of young, high, hot, thin, oceanic lithosphere would be mechanically easier than the detachment and overthrusting of older, low, cool, thick, oceanic crust and mantle. Third, the common direct association of ophiolites with rocks having volcanic arc affinities (andesitic flysch, andesites, tonalites, granodiorites) favors a rear-arc basin site for their origin.

Changes in basalt production through time

One important question is whether (and if so by how much) the amount of volcanism on Earth has fluctuated with time (see section 7.4.1). This question is fraught with difficulty. Generally, smaller proportions of older rocks are preserved so that the occurrence of volcanic rocks per unit time could be expected to be less; however, long ago, heat generation was greater so that volcanism may have been greater and these effects may partly cancel. Oceanic rocks are destroyed by subduction and obduction so that the record is necessarily incomplete and, generally, only arc and continental rocks are preserved. In spite of these difficulties, some workers have discerned episodicity in igneous activity, especially in the Precambrian. The crude technique of plotting histograms of numbers of isotopic ages, which was popular twenty years ago, has fallen into disfavor with the recognition of the complexity of the record (see Chapter 7). A particularly significant observation is that the continental areas representative of a particular time interval become comparatively much smaller for older rocks. For example, the Superior Province, which is the largest single area of Archean rocks on Earth (about half the world's total area of Archean not involved in orogenic reworking), amounts to only 1% of the present continental area. Several authors have taken a more sophisticated approach to episodicity by considering the isotopic compositions of Nd, Sr and Pb (e.g., McCulloch and Wasserberg, 1978). They have inferred that many Precambrian continental rocks became isolated from the main mantle reservoir during an interval of about 200 m.y., roughly 2.7–2.5 b.y. ago. This is a most interesting and unexpected result.

Turcotte and Burke (1978) applied an indirect method to the estimation of volcanic fluctuation with time during the Phanerozoic. Realizing that sea level responds to the volume of the mid-ocean ridges, they inferred that the times when the continents were most flooded were times when ridges were most active. Calculating, crudely, the proportions of heat escaping through conduction and ocean floor aging, they were able to estimate that nearly twice as much heat was escaping the Earth through the cooling of aging ocean floor during the late Cretaceous episode of continental flooding as is escaping by this means at present. In order to keep the Earth at roughly the same volume, these authors inferred that plate consumption was also at a peak during the late Cretaceous—a result consistent with the familiar high concentration of circum-Pacific batholithic emplacements during the Late Cretaceous and with the Late Cretaceous concentration of emplacements of Tethyan ophiolites.

Tectonics of basaltic eruptions at converging margins in the past

The igneous record of old convergent boundaries lies mainly in the intrusive bodies revealed by erosion in the cores of ancient mountain belts. These bodies, often batholithic in scale, have predominantly calc-alkaline compositional trends. For this reason, as well as reasons based upon structural considerations, they are commonly interpreted as representing the roots of volcanic arcs. Basalts, commonly as basaltic dikes, occur in small quantities in association with the calc-alkaline rocks of mountain-belt cores, but because of intense tectonism it is rarely possible to assign an individual occurrence to an unambiguous structural environment. Thus, basalts associated with extension (for example, in a pull-apart basin) in an overall convergent regime are hard to separate from basalts erupted from arc-volcanoes where both have been involved in subsequent deformation. Because of this kind of complexity, basaltic rocks representative of this environment have been studied relatively little, as yet.

Plate convergence eventually leads to continental collisions. These, in recent examples (Fig. 6.2.10), have been shown to give rise to dominantly calc-alkaline volcanics very similar to those found in Andean arcs (see section 6.2.1; Burke *et al.*, 1974; Kidd, 1975; Şengör and Kidd, 1979). Older examples of such calc-alkaline, K-rich volcanics and/or their subjacent post-kinematic granites are well known in collisional orogens up to about 2 b.y. old. Examples include the granites and other plutons of the Appalachian-Caledonian belt (Dewey and Kidd, 1974), the post-kinematic granites and local rhyolites of the Pan-African, the rhyolites of the St. Francois Mountains (Bickford and Mose, 1975) related to the collision-induced Elsonian orogeny, and the volcanics of the Bear Province (Hoffman, 1980) related to collision about 1.8 b.y. ago at the end of the Wilson cycle on the site of the Coronation orogen. Pieces of continental crust older than 2 b.y. are either too deeply uplifted and eroded, or are too small, or are too poorly known to allow us to distinguish collisional from arc-related magmatic products with any confidence; however, we see no reason to suppose that they are absent. As with the more recent examples (described in section 6.2.1), these magmas are conspicuously characterized by the dominance of intermediate and silicic rocks. Basalts are very subordinate and may even be absent in many areas.

Tectonics of basaltic eruptions at transform margins in the past

Oceans older than 200 m.y. are not preserved, so volcanic activity associated with old oceanic transforms is unknown except as fragmentary pieces in obducted ophiolite complexes (Karson and Dewey, 1978). Evidence of past activity has to be sought within the continents. Perhaps because of the small volume of transform-related volcanics and their susceptibility to later tectonic disruption, no well-documented examples are known from older, inactive transform zones. For example, well-studied pull-apart basins formed on an extensive, large-displacement Carboniferous strike-slip fault zone (a probable old transform system) through the Canadian Maritime Province (Belt, 1969) contain only very minor examples of contemporaneous volcanics. The exposed portion of the possibly correlative (Wilson, 1962) Great Glen Fault system does not show any obvious pull-apart structures. A much older example (~1.8 b.y.) of a preserved piece of a large-displacement strike-slip fault, the McDonald Fault (near the Great Slave Lake), also has no known volcanics associated with its movement, despite having very thick accumulations of coarse clastic sediments (P. F. Hoffman, pers. comm., 1977) that probably were preserved by pull-apart tectonics.

6.3.2 Tectonics of non-plate margin basalt in the past

Tectonics of basaltic eruption at hot spots in the past

Ophiolites, representing old ocean floor, locally contain alkali basalts, some of which may represent hot spot volcanic rocks carried into convergent plate boundary zones and later involved in obduction. Thrust faults are widespread in obducted terrains so that the alkali-basaltic assemblages in ophiolites are generally tectonically isolated. Middle Mountain, among the Franciscan mélanges of California (Berkland, 1972), constitutes an obvious example of what may be an alkali-basalt seamount. In general, the main tectonic significance of alkali-basalt occurrences in ophiolites is that they indicate that the ancient ocean floor, like the modern ocean floor, was studded with hot spot volcanoes. However, no alkaline and tholeiitic rock association that is large enough to represent an ancient Hawaii or an ancient Iceland has yet been recognized in mountain belts.

Just as there are active hot spot volcanoes within continents today, as best seen in Africa (e.g., Thiessen *et al.*, 1979), so there have been similar rocks in continents throughout the latter half of Earth history. Some ancient hot spots, like those of today, were associated with rift systems but others were isolated. It is difficult to assign exact tectonic significance to ancient isolated continental hot spot occurrences. In some cases, an ancient hot spot may have been contemporaneous with rifting. For example, North America was the site of active rift formation about 1200 m.y. ago and numerous isolated intracontinental igneous rocks of about the same age may be analogues of the isolated hot spots (e.g., Cameroun, Ahaggar) of Africa today. Burke and Wilson (1972) have associated the unusual concentration of hot spots in Africa with its being at rest over the mantle convective circulation. It is intriguing to speculate as to whether other outbreaks of intracontinental igneous activity remote from contemporaneous plate margins may also have been associated with a continental standstill.

The most complete records of intracontinental hot spot igneous activity are found in the Canadian Shield and in Africa, where large areas of old Precambrian rock are exposed. These areas display a record of hot spot igneous activity over a longer time interval than do continents with more extensive Phanerozoic cover. Two main kinds of intracontinental hot spot igneous activity can be distinguished in Africa and the Canadian Shield: those in which mantle-derived products such as kimberlites, basalts and carbonatites occur alone, and those in which mantle-derived products are associated with products of partial melting of the overlying continental crust. In some cases the distinction is clear; for example, the Jurassic granites of Nigeria (Jacobson *et al.*, 1958) are an excellent example of the latter type of igneous activity's being dominated by silicic material, which from its isotopic and trace element composition can clearly be seen to have been derived from continental melting. In other cases, such as the anorthosite terrain of Labrador, there is no predominance of continental crust-derived material, but it is not clear whether all of the hot spot igneous rocks in this province are mantle-derived.

Rifts in the past

In section 6.2.2 on active rifting, examples of ancient rifts were given to illustrate the main argument of the section: "rifts (elongate depressions overlying places where the lithosphere has ruptured in extension) occur in a great many different tectonic environments." Characterization of past tectonic environments is obviously more difficult than characterization of active regimes, and becomes progressively more difficult farther back in the past because the geologic record is less complete. A second characteristic feature of rifts emphasized in section 6.2.2 is that, whatever the environments in which they originate, rifts evolve to show very similar structures, stratigraphy and igneous activity. The continents contain very numerous ancient rifts (Burke *et al.*, 1978), but it is not easy to determine in all of these localities the type of tectonic regime in which they originated. For example, it is difficult, because of lack of accuracy of age determinations, to tell whether the Keweenawan system and its associated rifts formed in a continental rupture event of African type, as Sawkins (1976) suggested, or in the later collisional event that is commonly called the Grenville Orogeny. Somewhat similar rifts of lesser age are associated with either environment. One exciting possibility for future determinations is that features of rift-related basaltic volcanism may help in distinguishing the particular tectonic environment in which the rift formed. At present this has not been achieved. Basaltic and other volcanism in ancient rifts is varied in quantity, style and composition, but no correlations between volcanism and locality have been established. As in modern rifts, intense volcanism occurs in localized areas within the rift, rather than throughout the rift.

Flood basalts in the latter half of Earth history

Well-preserved remnants of older flood basalt and/or sill complexes are seen back to 2.15 b.y. ago. Most remnants are found within old rifts and aulacogens, because the accidents of erosion through geological time have claimed any more extensive basalts lying outside them. The most prominent flood basalt remnants of Paleozoic or older age lying outside old rifts and of any significant extent are the 600-m.y.-old Antrim basalts of Northwest Australia, the 1100-m.y.-old Keweenawan lavas and sills north of Lake Superior, and the 1900-m.y.-old Kimberly Plateau basalts and sills, also in Northwest Australia. Examples of flood basalts are more common within old rifts. Prominent ones in North America include the rifts of Keweenawan age (about 1100 m.y.), which include the Coppermine River (Northwest Territories), Seal Lake (Labrador), Gardar (Southern Greenland), Keweenawan and its extension in the Mid-Continent rift and gravity high, and various aulacogens along the Cordilleran margin of the North American craton, including one exposed in part in the Grand Canyon. Another prominent rifting episode is that shown by the extensive tholeiitic sill complex that invades the initial rifting facies clastics and volcanics (arkoses, etc., of the Seward and part of the Attikamagen Formations) preserved in the fold and thrust belt of the Labrador Trough (Baragar, 1967), whose age is about 2.15 b.y.

Associated with this 2.15 b.y. (Fahrig and Wanless, 1963) rifting and ocean-opening event is an extensive dike swarm (Stevenson, 1968) that runs (in Archean basement) subparallel with the northern end of the Labrador Trough and then turns to run obliquely along the Cape Smith belt, which is the continuation of the Labrador Trough in northern Ungava. This dike swarm is the last relic of a flood basalt event; a very similar, although smaller, example is seen close to and subparallel with the western margin of the Appalachians in northern Newfoundland (Williams, 1967), where it (and nearby small relics of flood basalts) was associated (Bird and Dewey, 1970) with the early Cambrian or slightly older opening of the ocean which Tuzo Wilson called the Proto-Atlantic (1966), now termed the Appalachian ocean or Iapetus.

Dike swarms as relics of flood basalt and rifting events are common. The MacKenzie dikes, which extend a huge distance (2500 km) north-northwest across the western part of the Canadian shield are of the same age as the Keweenawan rifting and flood basalt event. There are no well-preserved segments of similar continental rifted margins which are dated prior to 2.15 b.y. ago—the opening age of the Labrador Trough and Coronation oceans. However, there are examples of extensive tholeiitic dike swarms and we infer from these that rifting events, like those recorded in younger rocks, took place. The Ameralik dikes, which are found in West Greenland (McGregor, 1973) and are at least 3.0 b.y. and perhaps as much as 3.6 b.y. old (Pankhurst *et al.*, 1973), are evidence of the oldest rifting event directly recorded on Earth.

6.4 TECTONICS OF BASALTIC VOLCANISM ON EARTH IN THE FIRST HALF OF EARTH HISTORY

6.4.1 Tectonics of Archean basaltic volcanism

Jakeš (1973), among others, has shown that the average composition of continental crust resembles that of island arcs. He suggested that continents have been made by a two-stage fractionation of mantle material, the first stage producing ocean floor igneous rocks at ridges, and the second stage producing island arc igneous rocks above sinking, inclined slabs of oceanic lithosphere. Continents form from arcs by collision following horizontal motion. In the first stage, partial melting of mantle material to produce basalt and leave depleted pyrolite involves only vertical movement; however, once these rocks have been differentiated and emplaced, horizontal motion comes into operation as the new lithosphere is carried away from the spreading ridge. Cooling and thickening of aging lithosphere in this process is important in the convective dissipation of heat generated within the Earth. Because the thickness and elevation of cooling lithosphere change as the square root of the lithosphere's age (Parker and Oldenburg, 1973), more heat is dissipated by fast-spreading ridges than by slow-spreading ridges.

Figure 6.4.1 shows that much more heat was being generated by radioactive decay in the Earth 2.5 b.y. ago than now (see section 9.5.1). Many authors (for example, Goodwin, 1973) have emphasized the close resemblance of the rocks formed in the Archean and preserved in greenstone belts to those formed at ridges and above sinking slabs today. It is evident (Jakeš, 1973) from this resemblance that the two-stage fractionation of mantle material at ridges and above descending slabs was operating then much as today. Because two or three times as much heat was being generated, and because there is no evidence of the operation of any heat-dissipating process peculiar to ancient times, we infer that ridge processes were operating two or three times as fast as today. This greater intensity of operation could have been achieved either by greater average spreading rates or by greater total ridge length, or by both. Assuming that the volume of the Earth was about the same as now, there would have been either a greater length of subduction zone to consume the extra ocean floor or subduction would have worked more rapidly. Since there is some evidence that there is a general average for subduction rates at about 6 cm/yr (with respect to one of the slowly moving reference frames), a substantially greater length of subduction zone seems more likely.

The picture that emerges is that of an ocean in Archean time with a great length of fast-spreading ridge and a large number of long trenches, the two processes producing, at a great rate, ocean floor and arc rocks very like those of today. There seems little likelihood that oceanic thermal gradients were much greater in those times. Partial melting of pyrolite to make basalt presumably took place at temperatures and pressures similar to those of today. The greater amount of heat coming from within the Earth was dissipated mainly by the faster convective removal of basalt and depleted pyrolite at the faster-spreading ridges. Ultramafic lavas appear to have been rather more common in the Archean than in later rocks and their occurrence perhaps indicates local melting of the ancient pyrolite at higher temperatures. Since even in the Archean the proportion of these ultramafic lavas within basaltic piles was low, it seems unlikely that thermal gradients at ridges were much steeper than now. We suggest, as did Moores (1973), that Archean oceanic crust is likely to have been somewhat thicker than present oceanic crust, both because the mantle was then likely to have been a little less depleted, and because a slightly higher percentage of partial melting occurred due to slightly higher thermal gradients. In contrast, the Archean oceanic lithosphere (and continental lithosphere) is likely to have been (on average) somewhat, but not very significantly, thinner than at present because of somewhat higher conductive thermal gradients in the lithosphere.

Because all oceans soon "self-destruct," it is impossible to make direct observation of extensive areas of old oceanic material and inferences about ancient ocean can only be made indirectly, or from the dismembered slivers of oceanic and arc rocks to be found in greenstone belts. In contrast, continents are "non-geodegradable" and once made, they persist, although they are subject to episodic internal fractionation (Dewey and Burke, 1973) due to continent-continent collision.

Hargraves (1976) presented a non-uniformitarian model of an Archean Earth with a continuous sialic cover about 8 km thick overlain by 2 km of water. Mantle beneath the sial was suggested to convect and partially melt to yield basalt that broke the sial and erupted to form greenstone belts. This model is quite unrealistic. Greenstone belts are not piles of undeformed deep-water basalt; the rocks they contain and the structures that affect them so closely resemble those formed later in Earth history that there is no reason for invoking tectonic environments peculiar to early times.

We also reject the idea that greenstone belts record early impact history as Dietz (1965) suggested and as others have repeated. This is because the high impact flux in the solar system declined before the oldest preserved terrestrial rocks were formed and because the dynamic processes at the Earth's surface have been so intense as to destroy any possible record of the high impact flux. The condition of the Canadian shield as a recorder of late impacts shows how trivial these effects are when compared with terrestrial tectonic processes.

The 1850 m.y.-old Sudbury structure (lopolith) is the only well-documented terrestrial example of infilling of a large meteorite crater by basaltic magma where, from the coincidence in age, the impact probably triggered internally-fed magmatism (Dietz, 1964; French, 1970). The Bushveld Complex (about 2000 m.y. old) may be another, perhaps due to simultaneous impact by several large bodies (Dietz, 1963b; Rhodes, 1975). No other comparable examples are known of these rather distinctive, layered noritic rocks in clear association with indications of impact and impact melts. Most well-documented sites of large impacts, for example the 214 m.y.-old Manicouagan structure (Phinney *et al.*, 1978), show no evidence of associated basaltic magmatism but only of directly impact-generated non-basaltic melts and breccias. The geologic record thus suggests that impact-triggered basaltic magmatism has been a minor process on the Earth since heavy impact bombardment ceased about 3.9 b.y. ago.

Archean continental geothermal gradients

Archean geothermal gradients have often been supposed to have been much steeper than those of today (Fyfe, 1973; Baer, 1977; Strong and Stevens, 1974; Windley, 1977; Lubimova, 1969). Because heat generation from the decay of radioactive nuclides in the Earth was then much greater (Lambert, 1976; Lee, 1967; see Chapter 9), extra heat, if it escaped by conduction, must have been carried along steeper thermal gradients. Measurements of terrestrial conductive thermal gradients are very difficult to obtain, and estimates of thermal gradients in the past are even harder to make. Estimates made from igneous and metamorphic rock occurrences are of limited value, both because igneous rocks occur at levels in the Earth shallower by unknown amounts than those at which they form, and because nearly all igneous and metamorphic rocks are formed at plate boundaries where thermal gradients are much steeper than they are within plates. In plate margin areas, advective and convective processes dominate, so the normal conductive gradient of the lithosphere can only be measured or estimated in areas away from plate boundaries. The rocks of the Superior Province (formed roughly in the interval 3100–2500 m.y. ago) were not subjected to steep regional geothermal gradients after their assembly into a continent by lateral accretion of island arcs (Burke *et al.*, 1976), and hence the extra Archean heat escaped the Earth in some way other than by conduction through the continent. It has been suggested that this escape was mainly by cooling of the ocean floor as a boundary conduction layer (Burke and Kidd, 1978).

The Superior Province (Goodwin, 1972) consists of greenstone belts, the rocks of which have been compared to rocks formed in various island arc and marginal basin environments, and of intervening gneiss belts that have been compared to more deeply eroded arcs or microcontinents (Burke *et al.*, 1976). Although the gneiss belts show great structural complexity and may include some high-grade metasediments, granodioritic and tonalitic compositions greatly predominate. The gneissic terrains can be regarded as having a granodiorite-tonalite average composition (Goodwin, 1972). Wise (1974a), by considering the control on continental thickness exercised by the depth of water in the ocean, argued that continents 2500 m.y. ago were about as thick as they are now. Seismic refraction measurements (Hall, 1971) in the Superior Province indicate a depth to the base of the crust of about 35 km. Rocks of granodioritic composition subjected to pressures appropriate to a depth of 35 km (about 10 kb) begin to melt at a temperature close to 700° C, leaving a dry and refractory residuum.

The products of melting, which would tend to rise buoyantly in the crust and thus become exposed by erosion at the surface, typically are granites of minimum-melting composition (roughly equal amounts of quartz, albite and potash feldspar). Such granites are abundant in many crystalline terrains (for example, the Grenville, Pan-African and Churchill provinces) and have been compared to the acid volcanics of Tibet produced by partial melting of thickened continent (Dewey and Burke, 1973). Minimum-melting granites are extremely rare in the Superior Province. We interpret this scarcity as an indication that the temperature 35 km down at the base of Superior Province crust in Archean time did not generally exceed 800° C. We reject the idea that minimum-melting granites may have occurred at higher levels now eroded from above the Superior Province, because the preservation of low-grade assemblages in the greenstone belts of the province precludes deep burial and because of the limited occurrence of granulitic rocks in the province. As the temperature at the base of the continental crust today is commonly estimated (Clark and Ringwood, 1964) to be about 500° C, we conclude that in Archean times a typical continental average geothermal gradient was less than 23° C/km compared with 17° C/km today (Fyfe, 1973). Because heat generation rates (see Chapter 9) in the Earth between 2500 m.y. and 3100 m.y. ago may have been as much as three times greater than at present (Lee, 1967), the possible difference of 30% in the slope of the continental average thermal gradient would have been insufficient to remove all the extra heat then being generated.

How might the extra heat have left the Earth? Heat leaves the Earth's surface today in three ways, each of which accounts for roughly one-third of the total: by conduction through the continents, by conduction through the ocean floor, and by making and aging oceanic lithosphere as a boundary conduction layer to mantle convection (see section 9.5.1). An accurate assessment of the proportions of heat removed by these three processes is not possible, but it is suggested (see Sclater *et al.*, 1980) that one-third may be too low an estimate for making and aging oceanic lithosphere. During the Archean, the two conduction processes could have accounted for more heat, as both the mantle and the overlying continental and oceanic crust would have contained more heat-generating nuclides, but the evidence from the Superior Province (most other Archean terranes are similar) shows that increased conduction of mantle-generated heat through the continents was not significant. Although conduction through the largely-unpreserved Archean ocean floor may have been greater than at present, another way of removing very much more heat from the Earth is by making and cooling extra ocean floor. This could have been done by having a greater total length of ridge in the Archean ocean, or by spreading and thus making ocean floor faster at ridges, or by both processes. Since the rate at which ocean floor cools decreases as a function of age, more heat is removed from the Earth if the average age at which the ocean floor is subducted is reduced.

Ultramafic komatiites (Brooks and Hart, 1974), rocks that may represent a much larger proportion of partial melting of the mantle than do normal basalts, are apparently almost restricted to the Archean and have sometimes been considered as evidence for very steep Archean thermal gradients (Strong and Stevens, 1974). These rocks occur as a small proportion of piles basalts very similar to mid-ocean ridge basalts. We follow the suggestion (McKenzie and Weiss, 1975) that ultramafic komatiites are melts from great depths that have risen diapirically to the surface and have escaped the near-surface equilibration that is represented by basalt. We associate this process with the more intense convective circulation that we believe characterized the Archean rather than with a steep conductive oceanic thermal gradient. Arndt (1977) showed that komatiite chemistry was consistent with partial melting of upwelling material which had already fractionated basalt. Only a percent or two additional melting is necessary. See section 1.2.2.

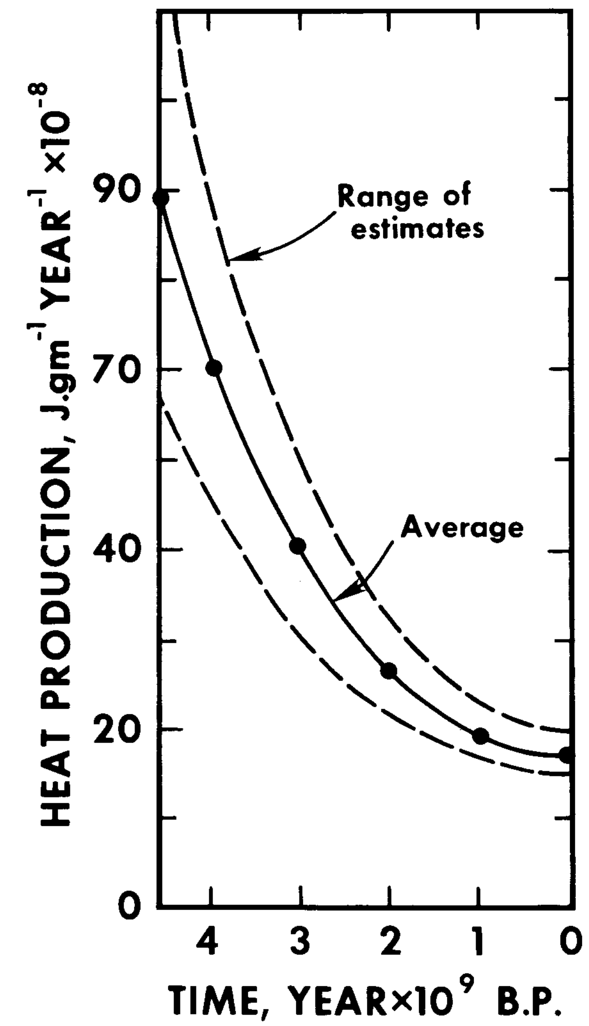
Archean tectonics: the style of the Superior Province

A basaltic dike swarm not younger than 3 b.y. and perhaps as old as 3.7 b.y. cuts the oldest rocks yet recognized on Earth (McGregor, 1973). This fact clearly shows that the upper lithosphere of that time behaved in a brittle manner under stresses similar to those responsible for younger dike swarms of which most in the Proterozoic and Phanerozoic are directly connected to rifting and the opening of oceans.

A general and sustained surface temperature of less than about 100° C since 3.4 b.y. ago is implied by the presence of thick sections of that age and of younger ages containing pillow lavas, although higher temperatures could have been sustained below deep oceanic waters and/or under a high atmospheric pressure. This observation, combined with reasonable estimates of heat generation (Fig. 6.4.1) and a likely distribution of the heat-producing elements, also requires that there was a rigid lithosphere of significant thickness (~20–40 km) by 3.4 b.y. ago. Because mantle heat production was at least twice the present rate at the end of the Archean, strain rates due to first-order lithospheric tectonics are unlikely to have been less than at present. Therefore, such tectonics must have operated within the brittle regime from the top of the lithosphere to a depth perhaps slightly less than, but essentially similar to, that of today. We therefore infer:

(1) that a rigid lithosphere has been present since at least 3.7 b.y. ago;
(2) that lithospheric relative movement has been by ocean-floor spreading, transform faulting, and subduction at least since this time, even though there probably was a significantly greater length of plate boundary, faster average relative plate motion, and many small "plates" prior to 2.5 b.y. ago; and
(3) that continental fragments behaved in a manner similar to those of later times, and that the compressive deformation seen within them is due to orogenic events localized in time and space due to subduction and/or collision-related processes, which may have involved the local and temporary occurrence of greatly elevated thermal gradients and large volumes of silicic magmas.

Sequences composed dominantly of mafic rocks over 10 km thick, all in low greenschist metamorphic facies



6.4.1 Heat production from radioactive decay of isotopes of U, Th, and K through geologic time. After Lee (1967).

(e.g., Glikson, 1972), whether composed of several thrust sheets or not, clearly demonstrate that equilibrium thermal gradients, at least in these areas, were not extreme in Archean times (not above 25° C/km).

Alkaline volcanic and/or plutonic rocks seem to be absent from rocks formed prior to 2.5 b.y. ago (with one exception noted below). This absence of alkaline magmas perhaps indicates that in environments now occupied by alkalic magmas (hot spots with or without associated rifting or spreading, in both continents and oceans), the percentage of mantle partial melt was relatively high (>10%), that perhaps this melting took place at shallower depths (<40 km), and may suggest that a somewhat thinner lithosphere was present prior to 2.5 b.y. ago.

No failed rift arms have been recognized prior to about 2.2 b.y. ago (2.4 b.y. ago if the southernmost part of the Huronian is in an aulacogen). We suggest that the faster pace of plate motion and the thinner lithosphere prior to 2.5 b.y. ago resulted in very few failures of rifts to evolve into oceans and, therefore, that there are no Archean aulacogens preserved in the small sample of essentially unmodified Archean crust remaining for inspection. Two types of Archean crustal terrain seem to be present. One is a "granitic," "gneissic" or "continental" terrain, which has been most thoroughly studied in West Greenland (Bridgwater *et al.*, 1974). The other kind is the greenstone/granodiorite or "oceanic" terrain. We consider the Superior Province the best example of such a terrain because its area greatly exceeds that of other greenstone/granodiorite terrains and because it has been so thoroughly studied. It is evident that the tectonic environment responsible for greenstone/granodiorite terrains such as the Superior must be compatible with plate tectonics because the Birrimian greenstone/granodiorite terrain of West Africa (~2.1–1.8 b.y.) is the same age and younger than the age of opening (~2.15 b.y.) of the ocean now represented by the Labrador Trough–Circum-Ungava–Nelson Front suture, which gives clear evidence of the operation of plate tectonics from this time onward (Burke and Dewey, 1972, 1973a).

Island arc complex assemblages and tectonics compared with greenstone/granodiorite terrains

Several authors (e.g., Engel, 1968; Folinsbee *et al.*, 1968; Anhaeusser *et al.*, 1968; Jakeš and Gill, 1970; Burke and Dewey, 1972) have put forward arguments suggesting that the granodiorite/greenstone terrains represent island arc rocks. Present-day island arcs and related surroundings contain five significantly different tectonic environments, all of which might be represented somewhere in the granodiorite/greenstone terrains if, as is likely, the analogy is valid. These environments are (Fig. 6.3.2): (a) remnant arc; (b) marginal basin, including oceanic crust formed by back-arc spreading (Karig, 1971a); (c) island arc; (d) arc/trench gap; and (e) trench mélange. Environments (a), (c) and (d) include oceanic crust on which the arc was built (Fig. 6.3.2), and (e) includes pieces of oceanic crust and overlying pelagic sediment in thrust slices and as tectonic blocks in mélange. Upon arc-arc or arc-microcontinent collision, the arc/trench gap and trench mélange are extremely likely, judging by their rarity in Phanerozoic orogenic belts, to be destroyed by being telescoped, sutured out, overthrust and eroded, perhaps because they are founded on essentially unmodified oceanic crust, which is readily subducted and/or obducted. Arcs involved in collisions tend to be uplifted and the calc-alkaline volcanic and volcaniclastic pile eroded off, which exposes the granodioritic plutonic terrain underneath. An example of this behavior includes the Ladakh "granite" (mostly granodiorite) in the Himalayas, which represents the roots of a volcanic arc that collided with the Asian continent in the Cretaceous. It is surrounded on both sides by narrow belts of ophiolitic rocks including, on the southern side, "flysch" derived by erosion from the arc (Gansser, 1964). Another example of this behavior is the Mesozoic batholithic belt of southern Chile, where the only evidence of the previously overlying volcanic rocks is now found in volcaniclastic sediments overlying the ophiolite complex floor of a Cretaceous marginal basin to the east (Dalziel *et al.*, 1974). This marginal basin opened by spreading behind the arc and closed by collision of the arc with South America, deforming the oceanic crust and overlying sediments in upright folds and steep tectonic slide zones (Dalziel *et al.*, 1974), and forming a greenstone belt (Tarney *et al.*, 1976). Even without collisions, volcanic arcs tend to end up being represented by their plutonic roots; for example, in the Sierra Nevada batholith complex of California. These examples illustrate that it is more likely that the volcanics and volcaniclastics preserved in geenstone belts are, where they are narrow anastomosing belts separated from one another by large areas of granodioritic material (such as in the Kenora area of the Superior Province), more likely to represent marginal basins, and possibly arc/trench gaps and "main" ocean sutures than they are to represent the island arc itself. However, where most of the outcrop area is greenstone with little granodiorite (as in the Abitibi area of the Superior Province), the arc volcanics, as might be predicted, seem to be well-preserved (Goodwin and Ridler, 1970). It must be emphasized that significant quantities of marginal basin and arc/trench gap rocks are also likely to be present in this and similar areas.

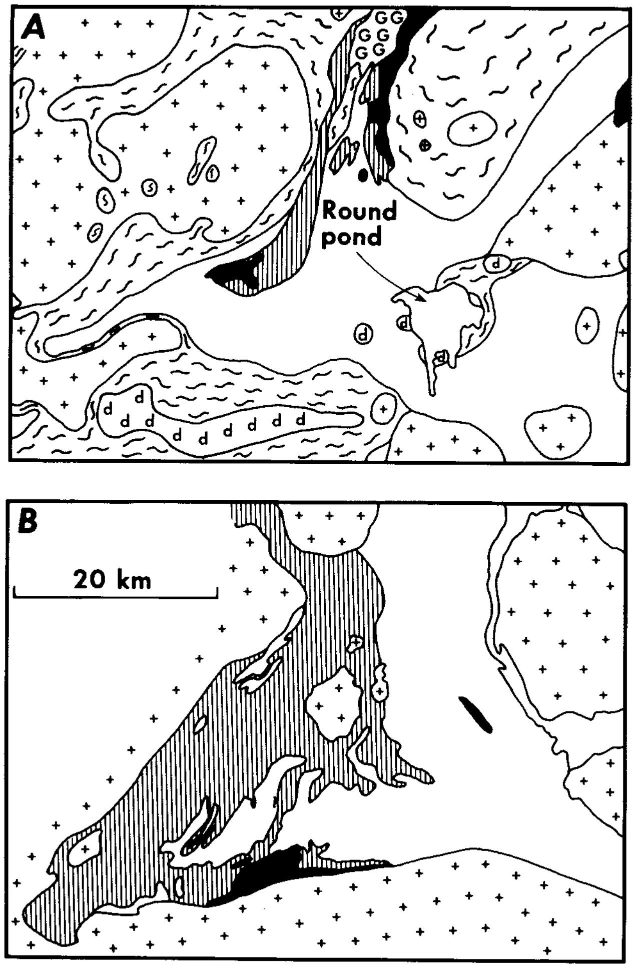
Although the bulk of lava flows in any one marginal basin—arc—arc/trench gap complex may be largely confined to the arc and to oceanic crust of various ages, they are not wholly restricted to the arc. Arc/trench gaps at any one time are defined by the trench and by the volcanic front which is controlled by the depth to the descending slab (Fig. 6.2.6). If the dip of the descending slab increases or decreases, volcanism will occur in what formerly was the arc/trench gap or marginal basin, respectively. Therefore, in an area representing an arc complex built over a significant time, the distinction between arc and arc/trench gap and between arc and marginal basin may not be clear-cut. Significant amounts of basaltic pillow lava are known within the mafic volcaniclastic sediment sequences overlying the oceanic crusts of two originally very narrow (< 50 km) fossil marginal basins in western Newfoundland (Upadhyay *et al.*, 1971; Kidd, 1977). These may represent a temporary shallowing of the dip of a subducted slab, or perhaps some other process peculiar to marginal basins. It is important that the ophiolite-complex pillow lavas (oceanic crust) of these two "fossil" marginal basins—but not the lavas in the sediments above—appear to be basaltic komatiites (Kidd, 1977). If, in addition to these processes and the opening and closing of marginal basins, arc complexes are constructed with continuous or episodic oceanward migration of the subduction zone (such as has been suggested for the Klamath Mountains in California; Fig. 6.3.2), great complexities of tectonic and magmatic chronology are likely, quite apart from additional complexities superimposed during arc-arc or arc-microcontinent collisions.

As well as granodioritic plutonic complexes representing the roots of arcs, granodioritic to granitic plutonism during and immediately following collision is probable, due to partial melting, caused by over thickening during collision, of arc or microcontinental crust, or of the underlying mantle forming the rest of the lithosphere (Dewey and Kidd, 1974). This process may have contributed a large proportion of the plutons in granodiorite/greenstone terrains, including most or all of the widely reported plutons that are synkinematic to some of the deformation sequences.

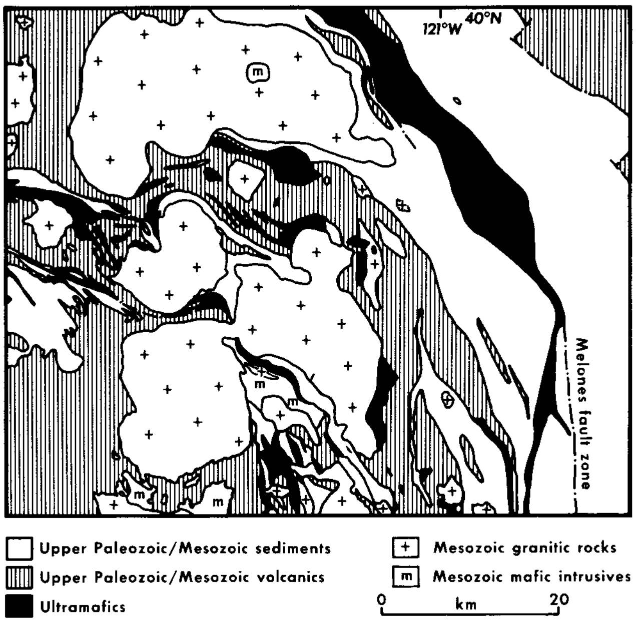
Two examples of alkaline volcanics and sills are reported from greenstone terrains. Both are in the Abitibi belt of the Superior Province: one near Kirkland Lake (Ridler, 1970), the other near Timmins (Goodwin, 1974). These may be island arc rocks, as suggested by Cooke and Moorhouse (1969) or, if there are more tectonic boundaries present than have presently been recognized, Archean seamounts tectonically incorporated into a trench complex (as is shown to be imminent in Fig. 6.3.2). Alternatively, it is possible that these two occurrences might be younger than the surrounding Archean rocks. Perhaps, judging by the largest presently observed, most of the hot spots were large in the Archean and erupted large quantities of flood basalt-type tholeiite relative to alkalic lavas. Lavas and sills generated in this way, and preserved from subduction, might explain the origin of at least some of the extensive basalts preserved in greenstone belts. Alternatively, efficient subduction of Archean oceanic lithosphere may have removed essentially all of the "hot spot" generated material and have preserved island arc edifices preferentially, which even today contain only rare occurrences of alkalic volcanics. Besides the small area of preserved Archean rocks, the small chance of preservation of alkalic volcanics may be a reason why they are so rare in such terranes.

Relatively small blocks, composed principally of quartzo-feldspathic rocks, occur within the dominantly greenstone/granodiorite terrain of the Superior Province, with the English River Gneiss Belt and the Quetico block being the best known examples (Goodwin, 1972). It has been reported that the rocks are largely "paragneiss," but further study seems to be supporting the view that they consist mostly of extremely deformed plutonic rocks, such as reported for the "gneissic" terrain of West Greenland by Bridgwater *et al.* (1974). The boundaries of the two blocks mentioned are reported as zones of intense faulting and mylonitization (Goodwin, 1972). In some areas, particularly on the northern side of the English River Belt in Manitoba (Wilson *et al.*, 1972), and to a lesser extent on the northern side of the Quetico block, tectonic slivers of serpentinite, and serpentinized ultramafic rock are clearly associated with these zones, giving them a similar character to oceanic suture zones of younger orogenic belts. Burke *et al.* (1976) suggested that these two blocks and others like them are microcontinents in terms of their present relationships to the granodiorite/greenstone terrains. Remnant arcs and inactive arc complexes can also be regarded for tectonic purposes as microcontinents with respect to active convergent plate boundaries and their active arcs. Such microcontinents may in some cases be identifiable in that they may contain a mixture of plutonic rocks of two greatly different ages, one reflecting its arc history, and a younger one reflecting its collision with an active arc that will only give ages near and including the younger age range. Such a relationship was suggested by Lawson (1913) to be common in the Superior greenstone/granodiorite terrane on the basis of clasts of granitic rock in greenstone belt conglomerates.

Lithologic associations very similar to those in greenstone/granodiorite terrains and deformed in a similar style can be found locally within Phanerozoic orogenic belts. One example is the Round Pond area of south-central Newfoundland (Fig. 6.4.2). This area, which includes a well-developed ophiolite complex, was deformed and intruded by silicic magmas during the Acadian continental collision, which was relatively weak in the salient of central Newfoundland (Dewey and Kidd, 1974). The "triangular syncline" morphology, the scale, and the lithological assemblage in the Round Pond area are very similar to those in Archean greenstone/granodiorite terrains, illustrated for comparison in Fig. 6.4.2 by the Bulawayan greenstone-belt (Amm, 1940). The steep metamorphic gradients shown by the narrow dynamothermal aureoles of the silicic plutons in the Round Pond area are a particular point of resemblance to the Archean terrains. An even better example is found in the northwestern Sierra Nevada of California (Fig. 6.4.3); the resemblance of this area to Archean greenstone/granodiorite terrains is remarkable. The ultramafic and gabbroic rocks in this area are almost all or perhaps entirely variably dismembered ophiolite complexes, including (just to the south of the area shown) one with well-preserved sheeted dikes (R. G. W. Kidd, pers. comm., 1975; Moores, 1975).



6.4.2 Geologic maps illustrating similar rock types, structural relationships and gross morphology in part of a Paleozoic orogenic belt and part of an Archean greenstone belt. **a.** Round Pond area, south-central Newfoundland (Williams, 1970). **b.** Bulawayan greenstone belt, Rhodesia (Amm, 1940). d—diorite; crosses—"granites"; wavy broken lines—high-grade metamorphics (probably dynamothermal aureoles around "granites"; black—ultramafic rocks; G—gabbros; vertical lines—mafic volcanics; white—sediments, including volcaniclastics. After Burke *et al.* (1976).



6.4.3 Map of pre-Cenozoic geology of part of northwestern Sierra Nevada, California, showing great resemblance in morphology and scale to Archean greenstone belts. Modified from Burnett and Jenkins (1962).

Tectonic boundaries and ductile high-strain zones are abundant (E. Nisbet, pers. comm., 1972). The ophiolite complexes and the mafic volcanics and volcaniclastics are probably at least partly of marginal basin origin.

It is important to recognize that the apparently swirly and plastic deformation and intrusion features seen in the map patterns of the Phanerozoic analogues cited for the Archean greenstone/granodiorite terrains are the sum of a complex series of events. In each case, these occurred during a significant time period in an area where the instantaneous situation at any time was either relative plate motion at sharply defined boundaries, or, during short-lived orogenic episodes, significant horizontal shortening with consequent upward motion of material to compensate. These episodes may have been due either to collision across a marginal basin or a "main" ocean, or to other (at present) poorly understood compressive orogenic episodes within and behind Andean and island arc complexes. There seems no compelling reason to suppose anything different for the Archean terrains.

An extensive gneissic terrain containing granulite-facies rocks forms the part of the Superior Province in the Ungava Peninsula. The radiometric ages (2.5–2.4 b.y.) and the nature of the rocks (Stevenson, 1968) show this to be an area of basement reactivation, which Burke *et al.* (1976) interpret as being due to convergence after a continental collision about 2.6 b.y. ago. This type of orogeny, where internal continental reactivation takes place (Burke and Dewey, 1973a), seems to have been relatively uncommon in Archean time, as shown by the large areas of granodiorite/greenstone terrain that have survived. Burke *et al.* (1976) suggested that it was the general incompleteness or weakly developed nature of collisions in the Archean that has led to widespread preservation of the greenstone belt material which is only locally preserved in younger orogenic belts.

The supposedly "vertical" tectonic style reported from greenstone belts seems to be based on the alleged control of deformation by the silicic plutons, the allegedly simple synclinal structures of the belts and, perhaps, on the subvertical elongation lineation commonly observed. The latter is a necessary consequence of horizontal shortening across a steep cleavage, because vertical displacement of plastically deforming material is usually easier than lateral displacement. The supposedly simple synclinal structure of greenstone belts is a myth (Ramsay, 1963). The steep inclination of cleavage and the overall steep structure of greenstone belts are quite compatible with collisional tectonics, being observed in Phanerozoic orogenic belts where incomplete collision has occurred (Dewey and Kidd, 1974). They also are found in very complete collisions in tightly sutured zones such as the Indus suture (Gansser, 1964). Although the Indus suture and others like it represent oceans driven out by low to moderately dipping subduction and thrusting, the suture zone itself becomes oversteepened in the later stages of the collisional process. This effect is also demonstrable in the Baie Verte Lineament (Kidd, 1977) on a scale more compatible with an individual greenstone belt. Ramsay (1963) has shown that the silicic diapirs do not significantly control the deformation in the Barberton greenstone belt. It is difficult to see how a strong regional cleavage and associated horizontal shortening could be impressed on relatively hard, cold volcanics and sediments by soft, ductile granitoid diapirs. An externally applied, horizontal compressive stress, such as is available during arc/arc or arc/microcontinent collision, is mechanically more reasonable, and is more compatible with the behavior of a planet on which occurred the major horizontal shortening clearly documented by Bridgwater *et al.* (1974) for the Archean of West Greenland.

Ophiolite complexes and Archaean oceanic crust and mantle

No examples of oceanic crust and mantle similar to fully developed ophiolite complexes have as yet been reported from greenstone belts. The suggestion (Moores,

1973) that Archean oceanic crust was thicker than present oceanic crust leads to the idea that possible examples of tectonic slices of Archean oceanic crust often may not include the depleted harzburgite-dunite that is a diagnostic and characteristic feature of their Phanerozoic counterparts. The somewhat higher heatflow and perhaps less-depleted mantle in Archean times may have resulted in a significantly greater thickness of essentially ultramafic cumulates and a proportionately lesser thickness of gabbro than is seen in Phanerozoic ophiolite complexes. As Moores (1973) pointed out, the stratiform anorthosites of the Early Precambrian (e.g., Fiskenaesset and the Limpopo) may also represent the upper part of the cumulate gabbro of the thick Archean oceanic crust. Many potential examples of Archean oceanic crust (regarding ultramafic cumulates as crust for brevity) are stated to be sills. A large number of Phanerozoic ophiolite complexes, now known to be variably dismembered and tectonically bounded slices of oceanic crust and mantle, were originally described as intrusive sills, despite the essential absence of significant metamorphic aureoles and the presence, usually overlooked, of narrow tectonic boundaries with large displacements. One example includes the Bay of Islands Complex (Smith, 1958). More dismembered examples, the Mings Bight and related Baie Verte Lineament Ophiolite Complexes (Bird *et al.*, 1971; Dewey and Bird, 1971; Kidd, 1977; Kidd *et al.*, 1978) were also described originally as concordant lenticular ultramafic and gabbro sills (Watson, 1943). Many examples of supposed sills which are very similar to the cumulate ultramafic and gabbroic parts of ophiolite complexes are known in Archean greenstone belts, and it has been suggested (Burke *et al.*, 1976) that many of them are good candidates for samples of Archean oceanic crust, perhaps including, in some instances, depleted noncumulate upper mantle as well.

These possible ocean floor fragments in greenstone belts with associated mafic volcanics and volcaniclastics closely resemble those dismembered ophiolites interpreted as being of marginal basin origin in Phanerozoic mountain belts (e.g., Bird *et al.*, 1971; Dalziel *et al.*, 1974). Burke *et al.* (1976) suggested that it is very likely that the contents of marginal basins compose a large proportion of Archean greenstone belts, and that there are two reasons why any Archean oceanic crust that may be identified will be more likely to represent marginal basin rather than main ocean crust. First, the Archean ocean with many scattered microcontinents and arcs appears likely to have had many more areas in which young ocean floor was close to arc and microcontinental sediment sources than in later times. Second, marginal basin ocean floor has a greater chance of being young (and hence thin and hot) near to a subduction-resistant arc or remnant arc (microcontinent) than does main ocean floor, and therefore has a greater chance of being obducted and preserved.

Summary of Archean tectonics

The characteristics of ancient ocean and continent outlined in this section have been inferred by assuming that rocks similar to those forming today were made by similar processes, and by allowing for the effects of faster heat generation in the past. Because much of the heat generated in the Earth today is dissipated in making oceanic lithosphere at spreading ridges, in aging it on ocean floors, in partly melting descending slabs of oceanic lithosphere below island arcs, and in emplacing the igneous products of this melting, Burke *et al.* (1976) have inferred that these processes operated more effectively during the Archean prior to 2.5 b.y. ago. The picture outlined by considering these intensive properties is of an Earth at the end of the Archean covered much as today with about one-third continent and two-thirds ocean, and the volume of water produced from the mantle concurrently with lithosphere being also similar to the present (Wise, 1974a). There are no minerals that will hold the water in the mantle at depths of much more than 100 km, so a hydrosphere containing much of the Earth's water must have existed from early times. Burke *et al.* (1976) have suggested that the length of plate boundary was, however, much greater and that this explains many of the differences. The extensive characters of exposed Archean terrains are consistent with this picture and permit its refinement. It is inferred that the environments of basaltic volcanism represented in Archean rocks were not very different from those of today.

6.5 TECTONICS OF BASALTIC VOLCANISM ON OTHER TERRESTRIAL PLANETS

Mars, Mercury and the Earth's moon share the property of having at present a stable global lithosphere. This is in contrast to Io, where lithospheric mobility is so great that all lithosphere is probably recycled frequently (on a time scale perhaps less than 10^7 years), and to the less extreme, plate-structured Earth. On Earth, oceanic lithosphere is fully recycled on a time scale of about 10^8 years, but most material that accreted