6. Tectonics of basaltic volcanism

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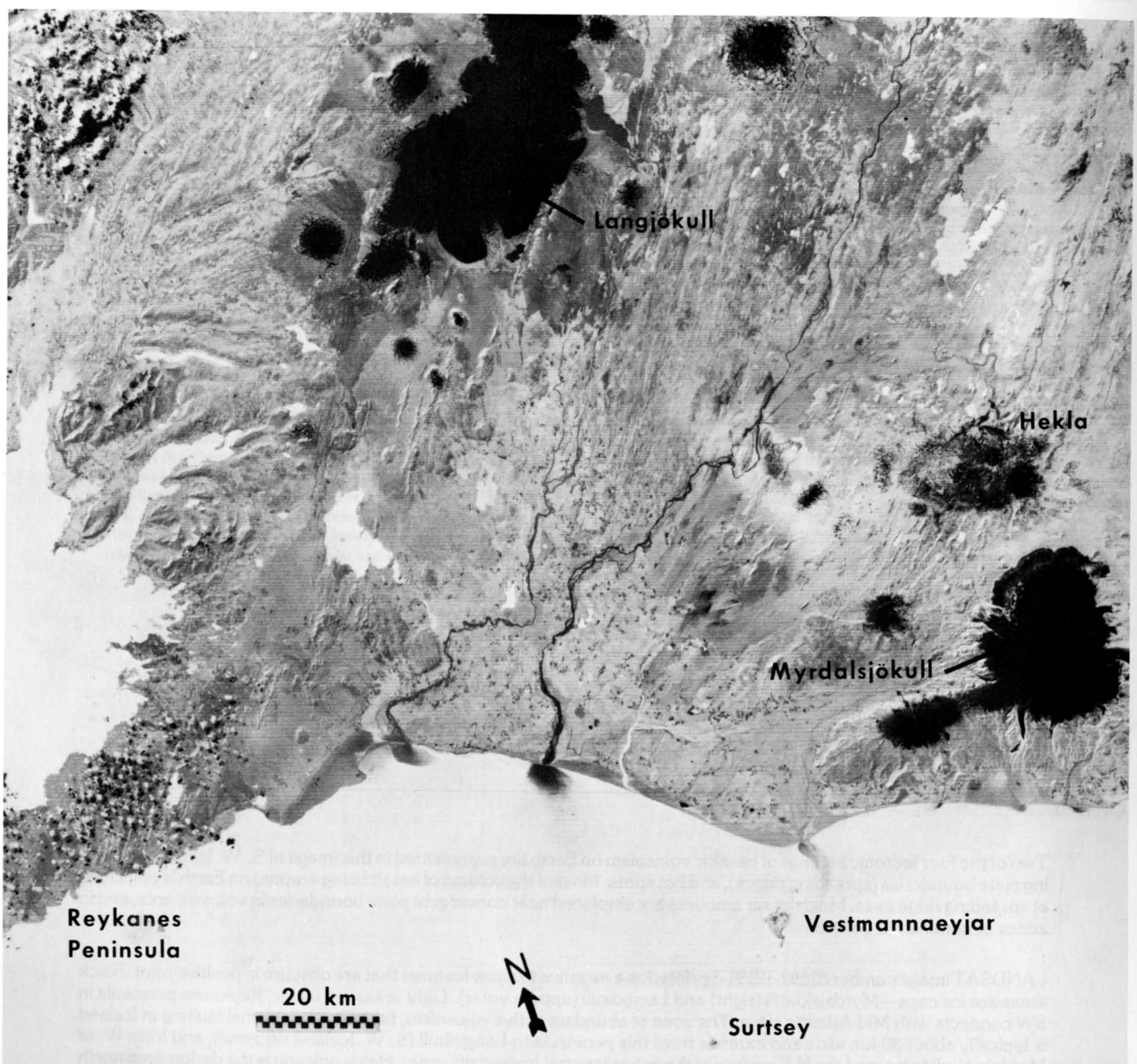


Overleaf:

Two of the four tectonic settings of basaltic volcanism on Earth are represented in this image of S. W. Iceland: accreting plate boundaries (spreading ridges), and hot spots. Most of the volume of basalt being erupted on Earth is emplaced at spreading ridge axes. Much lesser amounts are emplaced near convergent plate boundaries in volcanic arcs, and in zones of continental collision.

LANDSAT image number 81392-12191-5 printed as a *negative* to show features that are obscure in positive print. Black areas are ice caps—Myrdalsjökull (right) and Langjökull (upper center). Light areas are water. Reykanes peninsula in SW connects with Mid-Atlantic ridge. The zone of abundant active volcanism, fissuring and normal faulting in Iceland is typically about 30-km wide and extends from this peninsula to Langjökull (S. W. Iceland rift zone), and from W. of Myrdalsjökull to beyond the N.E. corner of the image (central Iceland rift zone). Hekla volcano is the darker area north of Myrdalsjökull, Surtsey is the small island near the lower edge of the print. Scale bar represents 20 km.

6.1 INTRODUCTION

Tectonics is the name given to the large-scale structural evolution of planetary lithospheres. On the Moon, Mercury, and to some extent Mars, nearly all of this structural evolution took place during a time when surface development was dominated by impacts. On Venus, the temporal relationship between lithospheric evolution and the impact history is enigmatic. For this reason, we have no unambiguous record of the tectonics of these planetary objects and emphasis in this chapter is necessarily very strongly on the Earth. A discussion of martian tectonics, especially that of the Tharsis region, is included here and comparisons are made with some terrestrial environments. Shorter reviews of the tectonics of the Moon, Mercury, and Venus are also presented.

Planetary tectonics, although essentially lithospheric, is the result of interaction between convecting mantle and overlying lithosphere, and very close links therefore exist between tectonics and the thermal histories of the planets. This is nowhere better seen than on the Earth where the plate tectonic process (Wilson, 1965a) plays an essential part in removing heat from the surface of the planet (Sclater and Francheteau, 1970; Turcotte and Burke, 1978).

In discussing the relationship of basaltic volcanism to tectonics, classical geological procedure is followed by considering first the presently active systems and then extending to the past the lessons learned from these. Known active volcanism in the solar system is restricted to the Earth and to Io; our discussion must start with the Earth because little is known yet about the volcanism on Io.

6.2 TECTONICS OF ACTIVE BASALTIC VOLCANISM

A revolution in understanding the lithospheric structure of the Earth was achieved when Wilson (1965a) pointed out that the lithosphere is broken into rigid plates that move toward, away from, or past each other. (The boundaries of these plates are shown in Fig. 6.2.14.) Tectonic and igneous activities are strongly concentrated at these boundaries. For students of basaltic volcanism it is convenient to consider basaltic activity at the three types of plate boundaries separately, and then to go on to consider the several varieties of non-plate-margin volcanism that together account for only a very small proportion of total terrestrial igneous activity. Active volcanism, possibly basaltic, is also occurring on Io, so its tectonics are briefly considered in this section. For a discussion of the possible mechanisms by which the plates are driven (all traceable to the decay of radioactive nuclides in the mantle) see section 9.5.1.

6.2.1 Tectonics of basaltic eruption at plate margins

Tectonics of basaltic eruption at accreting plate margins

Introduction. The majority of the basalt that is presently generated on Earth is emplaced along accreting plate boundaries (oceanic spreading centers). Menard (1967), Deffeyes (1970), and Dickinson and Luth (1971) have estimated the rates of production of basalt at spreading centers from the volumes produced since the late Mesozoic. The rates obtained are about 5–15 km^3/yr, which is only about one or two orders of magnitude greater than the present rate of magma production in the single most active non-plate-margin volcano on Earth, Hawaii (Lipman, pers. comm., 1978). The generally constant thickness of the (basaltic) oceanic crust shows that magma production rates at spreading ridges generally are remarkably steady, in contrast to intraplate-type volcanic areas. The latter, as shown by Hawaii and the Hawaiian-Emperor chain, may produce basaltic magma at a relatively high rate for the small area involved, but only for short times (on the order of 1 m.y. or less). Flood basalt areas, which are clearly associated with the largest hotspots and particularly with continental rifting (see section 6.3.2), probably have the highest magma production rate known, but only for the very short overall lifetime of these events (Baksi and Watkins, 1973). Thus, averaged over tens of millions of years, the volume of basalt produced at spreading ridges greatly exceeds that in other environments.

The recognition of the extent and continuity of the oceanic ridge system (Ewing and Heezen, 1956; Heezen, 1960), and the explanation of symmetrical magnetic anomalies by the seafloor spreading process (Vine and Matthews, 1963; Morley and Larochelle, pers. comm., 1962) are results of extensive oceanographic surveys. Heezen (1960) recognized that much of the ridge system in the Atlantic and Indian Oceans has an axial rift valley. The ridge in the Pacific Ocean was shown mostly not to have such a rift (Menard, 1964). The ridges, viewed on the broadest scale, are about 2000 km across; the axial part is generally about 2.5 km below sea level and the flanks subside to 5 km depth at a distance of about 1000 km from the axis. The subsidence of the ocean floor is clearly related to the cooling and thickening of the lithosphere away from the axis, and follows a quantitatively predictable path proportional to the square root of time, at least out to a crustal age of about 80 m.y. (Parker and Oldenberg, 1973; see Chapter 9).

The more detailed topography (see Chapter 5) is characterized by elongate ridges and valleys parallel to the axis of the ridge. This topography is more pronounced on ridges that possess an axial rift valley. It becomes progressively blanketed and obscured by sedimentation away from the ridge axis. The axial region of spreading ridges coincides with a narrow zone of shallow seismicity (Ewing and Heezen, 1956; Barazangi and Dorman, 1969; Sykes, 1967). Basaltic volcanism is also restricted to a zone a few kilometers or less in width along the ridge axis, and is almost wholly restricted to the inner floor of the rift valley when one is present (ARCYANA, 1975; Ballard *et al.*, 1975; Macdonald, 1977). The interaction of tectonics and volcanism in this axial zone is the key to understanding the formation of the oceanic crust and upper mantle and hence to understanding the occurrence of most of the basalt erupted today in the solar system.

Tectonics of spreading ridge axes. The presence or absence of an axial rift valley can be correlated to a large degree with spreading rate; above about 3 cm/yr half-spreading rate, an axial rift is not usually observed. Exceptions do occur: the Galapagos spreading ridge possesses a shallow rift where it is spreading at about 3.5 cm/yr half-rate (Hey, 1977), and the well-known Reykjanes ridge, spreading at about 1.5 cm/yr half-rate, does not have a rift. Both of these ridges are near sites of anomalous basaltic volcanism (Galapagos, Iceland) that have been termed "hot spots" (see section 6.2.4). Although there are likely to be differences between oceanic crust made at rifted and non-rifted (and hence slow- and fast-spreading) ridges, the tectonic processes involved are exclusively extensional and there must be more similarity than difference between crust formed in the two situations. Much detailed investigation of rifted spreading centers has been accomplished (see section 5.2.8), whereas there is much less information on nonrifted, fast-spreading ridges. Most of the detailed information on slow-spreading ridges has been obtained by deep-towed instrument packages and manned submersibles (Atwater and Mudie, 1968, 1973; ARCYANA, 1975; Ballard and Van Andel, 1977; Bryan and Moore, 1977; Macdonald and Luyendyk, 1977; CAYTROUGH, 1979). In addition, in two places (Iceland and the Afar region) the axis of the spreading ridge is emergent. Although, for that very reason, these two places are not typical, much can be learned from them about spreading ridge tectonics (Walker, 1975; Palmason, 1973; Palmason and Saemondsson, 1974; Needham *et al.*, 1976).

The rift valley, which typically is found in axial regions of slow spreading ridges, is a 30- to 35-km-wide, 1- to 2-km-deep trough, the inward-facing walls of which consist of a series of steps. The steps are comprised of a series of steeply dipping inward-facing scarps having relief of tens of meters to a few hundred meters. Some of the steps show clear outward tilt. The deepest part of the rift valley is commonly characterized by a relatively flat region 2–10 km wide called the inner floor (Needham and Francheteau, 1974). Most of the volcanic activity occurs in the inner floor, and lava flows build lenticular piles that are up to 200 m high in the FAMOUS area (Bellaiche *et al.*, 1974; ARCYANA, 1975; Ballard *et al.*, 1975). In Iceland, the zone of surface magmatism is largely between 30 and 40 km wide; most eruptions occur through fissures 2–10 km long and a few meters wide that are not concentrated in any one part of the rift. Dikes therefore feed subaerial flows that travel laterally for large distances (>5 km); this contrasts markedly with the situation in the areas (FAMOUS, Galapagos ridge) that have been observed by submersibles, although it is possible that some flows travel larger distances laterally in the submarine environment than have been observed. The observation of chemical zonation in basalts (all of very young age) in the FAMOUS area (Ballard *et al.*, 1975; Bryan and Moore, 1977) implies that the flows seen at the present surface apparently have not moved more than a few hundred meters from their points of extrusion; this observation also implies a zoned magma chamber below. The FAMOUS study demonstrated that few young extrusive rocks occur beyond the first major rift wall fault scarps. In Afar (Ghoubbat-Assal rift), extrusion also is essentially limited to the rift inner floor (4–10 km wide). Flows tend to be dammed by the rift walls and/or to flow along strike (Needham *et al.*, 1976). The locus of active volcanism and thus of dike intrusion appears to migrate within a 2-km-wide axial region of the Galapagos spreading ridge on a time scale of thousands of years (Van Andel and Ballard, 1979).

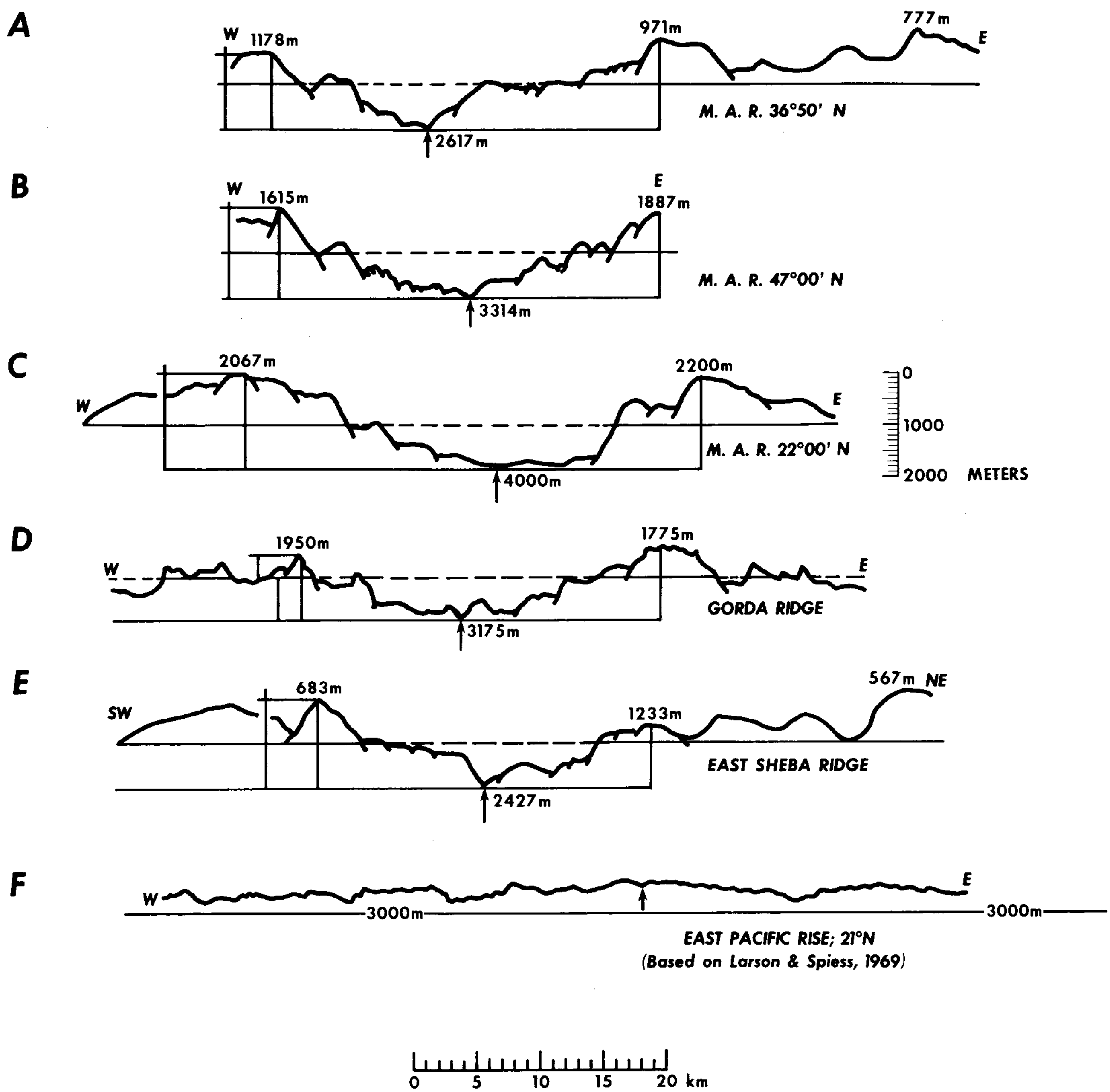
The lateral extent of flows, and therefore the width and constancy of position of the zone of extrusion, must affect the magnetic signature of the oceanic crust. Most of the magnetic signal comes from the upper 500 m or so of the crust (e.g., Lowrie, 1979; Schouten and Denham, 1979; Hall and Robinson, 1979), which appears to be largely composed of extrusive lavas. Atwater and Mudie (1973) inferred from the magnetic data that the zone of extrusion on rifted ridges is not more than a few kilometers wide; Larson and Spiess (1969) inferred by the same method that the zone of extrusion on a segment of the fast-spreading East Pacific Rise crest was no more than 300 meters across; this result has yet to be confirmed by extensive direct observations, but CYAMEX (1978) report a 0.4–1.0 km wide extrusion zone at 21°N.

The rift valley is usually grossly symmetrical with respect to the inner floor, but in some places (e.g., FAMOUS area, Fig. 6.2.1.a) there is moderate asymmetry. The whole structure is best explained as a series of not-necessarily-centered, nested grabens, the steep scarps of the walls being normal faults. Normal faulting takes place over a width of 10–15 km (Needham *et al.*, 1976; Lépine and Ruegg, 1976; Klein *et al.*, 1973). In most profiles, it is inferred that the outward tilts of the steps increase away from the axis of the valley together with a decrease in the dip of the scarps (Fig. 6.2.1). The hummocky terrain, with the peaks and troughs that constitute the rift mountains province, stands at a remarkably uniform elevation above the inner floor (Needham and Francheteau, 1974; Fig. 6.2.1). The floor of the rift valley deepens along strike toward the rift valley-transform fault intersection (Schroeder, 1977); for example, a 500-m deepening near fracture zone A has been observed in the FAMOUS area (Renard *et al.*, 1975).

The two mechanisms by which crustal extension occurs are fissuring and conjugate normal faulting (e.g., White and Stroup, 1979). In the FAMOUS area, Iceland, and Afar, the two processes are equally active in the inner floor of the rift, but normal faulting becomes predominant in the outer part of the rift floor and in the rift walls. One of the conjugate sets of faults becomes predominant in the rift walls but neither set is dominant in the floor of the rift (White and Stroup, 1979).

Thicknesses and layering of the oceanic crust are obtained from seismic refraction studies. These studies (Raitt, 1963; Drake and Nafe, 1968; Ludwig *et al.*, 1970; Shor *et al.*, 1970) indicate that a two-layer crust having a total and rather constant thickness of around 6.5 km or slightly more is commonly present. These seismic studies provide ambiguous information on the lithologies composing the crust; dredging, deep-tow camera observations and comparison with ophiolites better constrain the possible constituents of the oceanic crust. The upper layer, usually termed Layer 2 (about 2-km thick), appears to consist of basaltic lavas, mostly pillowed, on the present surface of the oceanic crust. The deeper parts of Layer 2 are likely to consist of fine-grained basaltic rocks, which may be either intrusive or extrusive. The lower crustal layer, termed Layer 3 (about 4.5 km thick), is generally considered to be mostly gabbros (e.g., Fox *et al.*, 1973); this overlies mantle with velocities appropriate for fresh peridotite. Layer 2 can be divided near the ridge axis into two portions (2A and 2B; Houtz and Ewing, 1976). Fox and DeLong (1976) suggest that this division, which dies out farther from the ridge axis, is due to highly fractured and rubbly basalts in the upper part near the ridge axis, that become consolidated farther away by cementation in the cracks. In some places in the Pacific, Layer 3 has been found to have a lower division, termed Layer 3B (Maynard, 1970; Sutton *et al.*, 1971; Snydsman *et al.*, 1973). This may consist of slightly serpentinized peridotite (Bottinga and Allègre, 1973) or of pyroxene-rich cumulate gabbroic rocks (Fox *et al.*, 1973). Two studies have detected this higher velocity, lower part of Layer 3 in the Atlantic, in crust formed at a slow-spreading ridge (Fowler, 1976; Steinmetz *et al.*, 1977). The extent of its development is not known in either the Atlantic or Pacific oceans, and earlier suggestions that it characterizes crust generated at fast-spreading ridges are not well-established.

Refraction studies have been conducted near the axes of rifted ridges at several locations. On the Mid-Atlantic Ridge at 45°N, Keen and Tramontini (1970) have results only for crust of age 2–6 m.y. (western rift mountains). The Reykjanes Ridge (Talwani *et al.*, 1971) is anomalous in depth and morphology. In the FAMOUS area, the refraction data point to the presence of Layer 2 at the axis as well as a Layer 3 (Whitmarsh, 1973, 1975; Poehls, 1974; Fowler, 1976). The depth of the mantle is on the order of 5–10 km. A point of dispute is the velocity of the mantle, either normal (Whitmarsh, 1973; Poehls, 1974) or reduced (Fowler and Matthews, 1974; Fowler, 1976). The basement layer (Layer 2) has a constant thickness away from the axis as shown by Houtz and Ewing (1976). Depth of the mantle is thus approximately 5 km below the ocean floor at least as close as 10 km away from the axis. Similar results have been obtained from precise refrac-



6.2.1 Profiles of slow-spreading, rifted (**a**, **b**, **c**, **d**, **e**) and fast-spreading, nonrifted (**f**) ridges. Arrows indicate axes; M. A. R. = Mid-Atlantic Ridge.

tion surveys in the axial region of the Tadjourah-Ghoubbet-Asal Rift in Afar (Lépine *et al.*, 1972; Ruegg, 1975).

The work of Francis and Porter (1973) at 45° N in the Atlantic and of Whitmarsh (1975) in the FAMOUS area suggests the presence of a magma chamber at a depth of about 3–5 km. Seismic refraction and multichannel seismic reflection investigations of the East Pacific Rise crest indicate the presence of a low velocity zone, interpreted to represent the top of a shallow level magma chamber, at a depth of 2–3 km (Orcutt *et al.*, 1976; Rosendahl *et al.*, 1976; Reid *et al.*, 1977; Herron *et al.*, 1978). This is also the estimate of J. C. Ruegg (pers. comm., 1976) for the region of Tadjourah. There is evidence for strong attenuation of P and S waves below the axis (Molnar and Oliver, 1969; Solomon, 1973; Solomon and Julian, 1974; Ruegg, 1974; Pontoise *et al.*, 1976). Steinmetz *et al.* (1976) and Sapin (1975) show that at 40°N in the Atlantic, the zone of strong P wave attenuation is about 50 km wide, with sharp lateral boundaries and an upper surface deeper than 6 km below the seafloor. Thus, energy is transmitted across the ridge in the uppermost lithosphere but not deeper down. This is also true for the FAMOUS area, where shear waves propagate across the axis (Fowler, 1976). This effect was not seen with the approach of Molnar and Oliver (1969).

Weidner and Aki (1973) have shown that earthquakes occur at the axis at shallow depths (2 or 3 km). This estimate may be inaccurate, but J. C. Ruegg (pers. comm., 1976) has found, with a local seismometer network in Afar, that most seismic foci are in the crust (i.e., are at depths less than 6 km). This reflects the thickness of the layer in which deformation is brittle in the axial region.

Earthquakes in the rift valleys have low magnitudes (≤ 5.6; Francis, 1974). Earthquake swarms are common (Sykes, 1970; Tadjourah and Gulf of California), and high b values have been related to low stress (Thatcher and Brune, 1971; Francis, 1974). Teleseismic data for large earthquakes ($M \geq 5.5$) yield normal fault mechanisms (Sykes, 1967; and for Iceland, Klein *et al.*, 1973), which is in agreement with field observations where no evidence for compressive tectonics has ever been found. The accuracy of epicenter locations determined from teleseismic data is not sufficient to provide constraints on the width of the active seismic zone. The width of this zone can be estimated from local observations in Tadjourah (Ruegg, 1975) and at the Mid-Atlantic Ridge at 45°N (Francis and Porter, 1973), where seismicity is confined to the median valley (perhaps in the central 15 km of this valley), which is in agreement with tectonic observations.

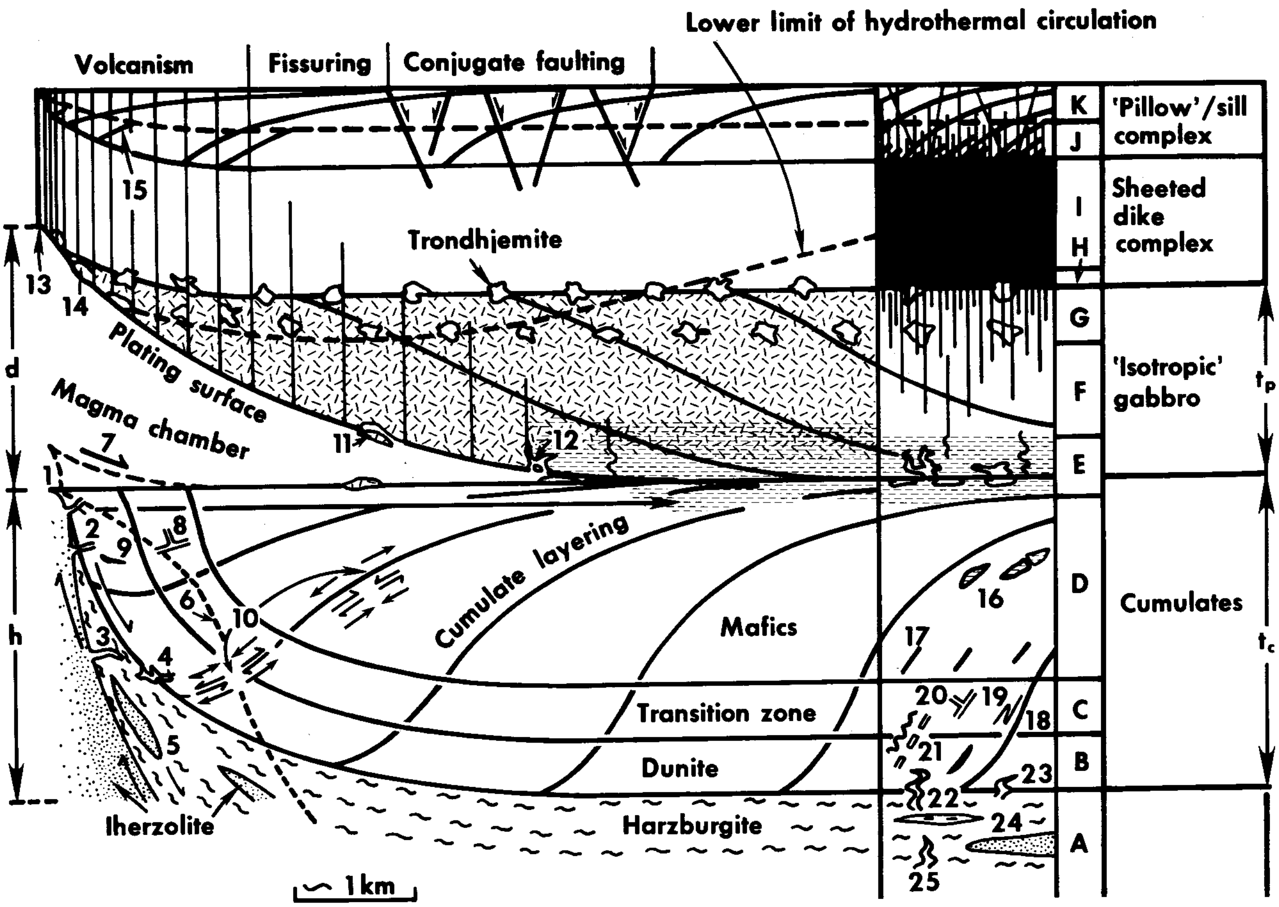
Evidence from ophiolite complexes. Models for the geometry of plate accretion at oceanic ridges have been developed by several authors who have adopted the view that ophiolite complexes (Penrose field conference participants, 1972) represent detached slices of oceanic crust and lithosphere (Dewey and Kidd, 1977; Kidd, 1977; Church and Riccio, 1974; Dewey *et al.*, 1973; Greenbaum, 1972; Moores and Vine, 1971). Numerous 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 and that, although the thickness of individual ophiolite complex units may vary considerably, the range of thickness variation is comparable with layer thickness variations recorded in the oceans (Coleman, 1971; Dewey and Bird, 1971; Moores and Jackson, 1974).

The general sequence of rock types in ophiolite complexes consists (Fig. 6.2.2), from base to top, of the following units: harzburgite-minor dunite-rare lherzolite, with metamorphic textures, and a complex and pervasive high-temperature deformation history; cumulate dunite below a cumulate transition sequence of dunite, feldspathic dunite, wehrlite, feldspathic wehrlite, clinopyroxenite, and gabbro-containing zones affected by penetrative high-temperature deformation; layered cumulate gabbro, clinopyroxenite, and anorthositic gabbro; more isotropic and homogeneous, somewhat leucocratic gabbro with irregular pods and veins of trondhjemite near the top and with diabase dikes increasing in abundance toward the top; a sheeted diabase dike complex with minor trondhjemite bodies near the base; and a pillow lava complex with an increasing percentage of dikes, and sometimes sills, toward the base. The following constraints imposed on the geometry of plate accretion are drawn mainly from observations on the Bay of Islands, Mings Bight and Betts Cove Ophiolite Complexes in western Newfoundland (Dewey and Kidd, 1977), but include those on other ophiolite complexes, particularly Troodos (Gass and Smewing, 1973; Greenbaum, 1972; Kidd and Cann, 1974; Moores and Vine, 1971) and Vourinos (Jackson *et al.*, 1975; Moores, 1969).

In most ophiolite complexes, from 500 m to 1 km of pillow lavas are developed, within which the percentage of dike increases downward until the base is defined as the level at which the lowermost screen of pillow lava is found, indicating that the lateral extension strain from fissuring increases downward. This shows that the complete pillow complex cannot be accumulated in batches of full thickness, but must grow to its full thickness some distance from the axis. Sills become common toward the base of the pillow complex in the Betts Cove Ophiolite, and in the Point Sal ophiolite, a 100-m-thick sill zone separates the homogeneous gabbro and diorite zone from the pillow complex (Hopson and Frano, 1977), indicating that sill injection may account for a part of the growth in thickness of the lower part of the pillow complex.

Sheeted-complex dike widths are commonly in the range of 20 cm to 1 m, which, for a half-spreading rate in the range of 1–10 cm/yr, indicates an average of one dike emplaced every few years to a few hundred years. Sheeted-complex dikes commonly preserve only one chilled margin with a preferred chilling direction, indicating that dikes are split to allow further dike injection and that most injection takes place over a very narrow zone away from which dikes are injected at a rapidly decreasing rate (Kidd, 1977). These dikes are notably and consistently aphyric. The fissures in which the dikes form must either be tapping magma without phenocrysts or tapping fluid segregating from a mush of crystals and interstitial fluid. It is possible that the crystal mush, where present, may ascend some way up from the bottom of any dike-filled fissure; this may partly account for the absence of sharply defined bases of individual dikes in ophiolite gabbros.

There is a strong contrast in grain size between dikes and gabbro, indicating a rapid cooling of gabbro before the emplacement of a particular dike. This evidence supports the idea (Anderson, 1972; Lister, 1972) of hydrothermal convective (rather than conductive)



6.2.2 Steady-state model for development of oceanic crust and upper mantle at an accreting plate margin. Top of pillow lavas taken as level datum. 1—magma supply axis to magma chamber. 2, 3, 4, 8—dikes intruded in residual harzburgite and lower cumulates (dunite, orthopyroxenites, websterites, clinopyroxenites, gabbros). 2, 8—undeformed. 3—folded. 4—strongly folded and disrupted. 5—lherzolite lens trapped in residual harzburgite. 6—zone undergoing penetrative ductile deformation due to shear between subsiding cumulates and rising melt plus residuum. 7—possible slumping due to non-steady-state rise of axial welt. 9—lens of chromitite from deformation of cumulate layers. 10—contrasting bending deformation due to subsidence of cumulates. 11—block stoped from magma chamber roof. 12—local intrusive gabbro. 13—axis of spreading (axial singularity); top of magma chamber. 14—trondhjemite bodies forming near axis. 15—sills injected into lower volcanics. 16–25—features seen in fully formed ophiolite complex due to spreading process; column at right side of diagram gives vertical section of fully assembled oceanic crust. 16—xenoliths from roof (11). 17—small-scale cumulate layering attitude near top of layered cumulates (oblique to large-scale layering). 18—folded layers in deformed cumulates. 19—undeformed dikes (gabbro, pyroxenite) in cumulates. 20—buckle folded layers. 21—boudinaged layers. 22, 23, 25—folded dikes (pyroxenite, dunite) in cumulates and harzburgite. 24—gabbro lenses in residual harzburgite (dikes or bodies of trapped magma). See Fig. 6.2.4 for terminology.

cooling of the upper part of the oceanic crust as it travels away from the axial singularity. The depth of the transition from sheeted-complex to gabbro and, therefore, the thickness of the 100% sheeted dike layer are most probably controlled by the lower limit of hydrothermal circulation at the axial singularity. The numerous anastomosing breccia zones observed within the sheeted-complex in some ophiolite complexes also suggest pervasive hydrothermal circulation (Williams and Malpas, 1972). In western Newfoundland ophiolite complexes, the transition from 100% sheeted dike complex to gabbro with less than 5% dikes takes place over a vertical thickness of about 250 m or less when viewed on a broad scale. On a finer scale, the transition is not homogeneous; at any level within the transition, zones a few tens of meters wide consisting mainly of dikes alternate laterally with similar zones consisting mainly of gabbro. In Mings Bight (Kidd *et al.*, 1978) one example of such a zone about 300 m wide consists, near its base, of 95% dikes penetrating about 200 m lower into gabbro than in nearby sections. Each zone of greater dike concentration up to a few hundred meters wide and deep was therefore a temporarily preferred site of dike injection.

In the homogeneous gabbro, the percentage of dikes decreases with depth. Dikes do not gradually increase in grain size to merge with gabbro. This means that the amount of lateral extension decreases from the base of the sheeted-complex to zero at the base of the homogeneous gabbro where the last dike dies out. Individual dikes cut solidified gabbro and are then underplated by gabbro as they move away from the axial singularity, just as dikes and flows in the pillow complex are progressively overplated by lava flows. Rocks of trondhjemitic type are found in the homogeneous gabbro, generally where it is cut by 5–20% dolerite dikes. A few occur within the base of the 100% sheeted dike layer. These bodies range from small individual tension gashes and veins through areas of net-vein breccia a few meters to tens of meters across and, exceptionally (Troodos), bodies as much as 1 km in original horizontal width. They are clearly associated with the sheeted-complex–gabbro transition and are part of the ophiolitic magma sequence, because many examples cut some dolerite dikes but are cut by others. One example in Troodos is so greatly dissected by dikes that it must have formed very near the axis of injection. Many of these bodies show clear intrusive relationships (net-vein breccias); their mineralogy, texture and relationships to adjacent rocks, particularly where they crosscut dolerite or contain dolerite or gabbro xenoliths, show that they are not hydrothermally altered gabbro, even though such alteration may occur locally.

The upper part of the overall gabbro layer, including that within the transition into the dike complex, is generally very homogeneous but locally shows diffuse layering a few centimeters or less thick. This homogeneous gabbro is somewhat richer in plagioclase when compared to gabbros of, for example, isolated thick tholeiitic sills and dikes. Plagioclase flotation in a magma chamber probably is not necessary to explain this slight enrichment in plagioclase, particularly because the clinopyroxene content is still large and grains are not interstitial. In addition, the presence of plagioclase-rich layers in the underlying cumulate gabbros suggests that plagioclase flotation is not important.

The thickness of homogeneous gabbro cut by diabase dikes is a poorly known parameter in well-preserved ophiolite complexes with thick gabbro units. The upper part with abundant dikes is clearly not less than 300 m thick, but firm data on the depth to which less abundant dolerite dikes occur are difficult to find. The Mings Bight Complex (Kidd *et al.*, 1978) contains diabase dikes through at least 750 m of gabbro. The Troodos Complex appears to contain diabase dikes nearly as far down as the base of the gabbroic layer; the thickness that this represents depends on the structural interpretation preferred for the central part of the complex. If it is a gentle dome (Wilson, 1959), then the depth is not more than about 1 km; if the western side of the central part is the steep limb of a monocline, it may be as much as 3 km. The Betts Cove Complex (Upadhyay *et al.*, 1971) has a very thin gabbro layer; where it is clearly cumulate, it contains no dolerite dikes, but where it is homogeneous and probably not cumulate, dikes are found throughout almost all of the 250 m of gabbro. The other complexes do not, at present, provide additional data.

It is important to emphasize that this discussion refers to dolerite dikes that do not cut the cumulate or the noncumulate ultramafic rocks. The fact that the dikes of the sheeted-complex are "rootless" means that they were fed from a magma chamber and that the gabbro they cut was plated onto the roof of that chamber. Later events which affect the already-consolidated oceanic crust, are local and may include hot-spot-type magmatism, arc volcano construction, and effects where a crustal block passes a ridge-segment end across a transform-fault–fracture-zone transition. In all cases, these dikes would be expected to cut cumulate gabbros and ultramafic rocks and the noncumulate ultramafic rocks. This can be seen in the Mings Bight Ophiolite Complex (Kidd *et al.*, 1978) where large, distinctively porphyritic, dolerite dikes cut all members of the ophiolite complex and feed large sills within the volcaniclastic sediment/pillow lava sequence that lies on the ophiolite complex proper.

Because of the difficulty of distinguishing plated roof gabbro from rapidly deposited, homogeneous, cumulate gabbro, the maximum thickness of cumulate gabbro in ophiolite complexes is poorly known. Not more than 500 m of layered cumulate gabbro has been proved to exist in well-preserved ophiolite complexes, although 2 km or more may exist in the Bay of Islands, Papua and perhaps the Oman; Betts Cove and Mings Bight contain not more than about 300 m.

The most reliably established section of cumulate ultramafic rocks comes from Vourinos (Jackson *et al.*, 1975); even though it is composite and crossed by faults, a minimum thickness of 1 km of ultramafic cumulates is present. This section shows cycles of deposition, which may indicate a batch or an intermittent supply of magma. Interpretation of the structure of the western margin of the Troodos massif as a steep monocline limb leads to a maximum thickness of 1.5 km for the ultramafic cumulates, including the lower 300 m of dunite. The Bay of Islands Complex probably contains a similar thickness, with the lower part also being cumulate dunite and feldspathic dunite. Although cumulate gabbros and transition-zone rocks and particularly cumulate dunite contain zones affected by penetrative high-temperature deformation, regular undisturbed cumulate layering in the gabbros and ultramafic cumulates demonstrates that ophiolite magma chambers have generally flat floors. Yet, in the Bay of Islands Complex, the orientation of cumulate layering relative to gross unit layering varies from roughly parallel at the base and top of the cumulate sequence to as much as 90° in the middle of the cumulate sequence. Although detailed cumulate layering cannot be traced laterally for long distances as it can in many layered mafic plutons, there is no evidence that crosscutting individual plutons can be defined within the gabbro and ultramafic cumulate layers. Therefore, a steady-state mechanism must be inferred for the development of cumulates and homogeneous gabbro.

The boundary between the cumulate and deformed noncumulate ultramafic rocks has been shown to be sharp and traceable for 1 km in the Vourinos complex; it is possible that this contact actually is between undeformed and deformed cumulates rather than between cumulates and noncumulates. In the Bay of Islands Complex, this boundary is somewhat obscured because the lower ultramafic cumulates are penetratively deformed, a situation that also seems to be the case in Troodos (George, 1978). The basal cumulates must therefore have been deposited on a noncumulate residuum which was sufficiently consolidated to preserve the gross cumulate layering but plastic enough to allow their deformation. As the spreading process pulls the lithosphere apart, the top of this harzburgite "basement" has to be formed at the magma supply axis immediately prior to deposition of cumulate material on it; it cannot be much older basement as some have previously suggested.

The lack of intrusive, crosscutting, mafic material in the residual harzburgite and overlying ultramafic cumulates of well-preserved ophiolite complexes shows that the extraction of partial melt at the supply axis is extremely efficient and must occur over a very narrow zone closer to the supply axis than the position at which the first cumulates are deposited. The occurrence of residual harzburgite immediately below cumulates and the existence of an at least 7-km thickness of harzburgite in part of the Bay of Islands Complex show that efficient melt extraction and residuum plating occur from the level of cumulate deposition and that they must continue progressively to at least 7 km below that level.

Models for plate accretion. Models dealing with plate accretion have addressed two fundamental classes of problems. One is that of the large-scale surface topography and tectonics; in particular, explaining the existence of a steady-state axial rift valley on slow-spreading ridges. The other concerns the generation of the various lithological units of the oceanic crust and upper mantle.

There are two broad sets of models that purport to explain axial rift valleys or the general topography of spreading axes. Kinematic models have been proposed by Deffeyes (1970) and by Anderson and Noltimier (1973), who suggested that the existence of a rift valley is a function of a budget between strain in the axial region and lava accumulation. The model requires that formation of the oceanic crust takes place in a region which is wider than the tectonic zone. This is not supported by the seismic results (Whitmarsh, 1973, 1975; Poehls, 1974) or by the field observations (ARCYANA, 1975; Ballard *et al.*, 1975; Ballard and Van Andel, 1977; Bryan and Moore, 1977) in the FAMOUS area.

Palmason's (1973) model provides an interesting kinematic analysis for Iceland. However, his boundary condition of a steady-state horizontal topography makes his approach not useful to the problem of rift valley formation. Stressing the importance of isostatic compensation, Daignieres *et al.* (1975) have proposed a similar numerical model based on Iceland data, where both tectonics and volcanics are considered as successions of discrete events.

The following mechanical models have been proposed: Piper and Gibson (1972) have suggested that the presence of rift valleys is a function of the width over which stress is applied through hydrostatic pressure in a magma chamber below the axis. The analysis yields a zone of horizontal compressive stress very close to the axis. For example, a magma chamber with a 2-km radius at a depth of 6 km produces compression over a region extending about 6 km away from the axis, beyond which active normal faulting takes place. Other authors (Lliboutry, 1976; Koide and Bhattacharji, 1975) have suggested a similar active role for the magma intrusion in the axial region.

Sleep (1969) has suggested a viscodynamic explanation for the origin of rift valleys in slow-spreading ridges. The central depression is explained by viscous head loss of the rising material in a vertical-walled cleft below the axial region. Recovery lifts up the crust and lithosphere in both walls of the rift valley. This idea has been the starting point for more elaborate models such as those of Osmaston (1971), Lachenbruch (1973), and Sleep and Rosendahl (1979).

Osmaston (1971) and Harrison (1974) have predicted reverse faulting on rift valley walls, which is in contradiction with observations (ARCYANA, 1975; Needham *et al.*, 1976). The viscodynamic model of Sleep (1969) and Sleep and Biehler (1970) raises some basic difficulties. Sleep (1969) proposed his model because he rejected the idea that rift valleys could be the result of graben formation on the basis of Vening Meinesz's (1950) analysis for grabens within continental crust. The latter is based upon a set of questionable boundary conditions: normal faults are proposed to cut through the whole rigid lithosphere, and no friction and zero bending movements are assumed on the fault plane. Also, Sleep (1969) attempted to model a rift valley as one single graben and found actual valleys to be too deep to represent a "normal" isostatic graben. In reality, a rift valley is a series of nested grabens.

Lachenbruch (1973) has made a quantitative treatment of Sleep's viscodynamic mechanism with the aim of explaining second-order structural features at the axes of fast- and slow-spreading ridges. In fact, only the gross features are modelled: valley for slow-spreading ridges and rise for fast-spreading ridges; the top crustal layer is ignored; the graben nature of the valley is not considered. The large number of independent parameters and the restrictive conditions which Lachenbruch (1973) set for the geometry of the conduit and the behavior (Newtonian viscous) of the rising material make his model difficult to test. Lachenbruch (1976) extended the model to conduits of triangular shape. Little attention was given to fitting observations. Sleep and Rosendahl (1979) have constructed steady-state viscous flow models of the topography of fast-spreading, slow-spreading, and 3 cm/yr ridges, using rheology inferred from detailed thermal models. If the material in the crustal magma chamber was assumed to have low density and low viscosity, reasonable fits to the observed topography were obtained. At 3 cm/yr spreading rate, the presence of a rift or axial high was sensitive to the crustal thickness. A greater crustal thickness resulted in a larger crustal magma chamber and thus an axial high as at fast-spreading ridges. A lower crustal thickness resulted in a small magma chamber as at slow-spreading ridges. Slow-spreading ridges with very thick crusts, such as the Reykjanes Ridge, are supposed to have axial highs for the same reason.

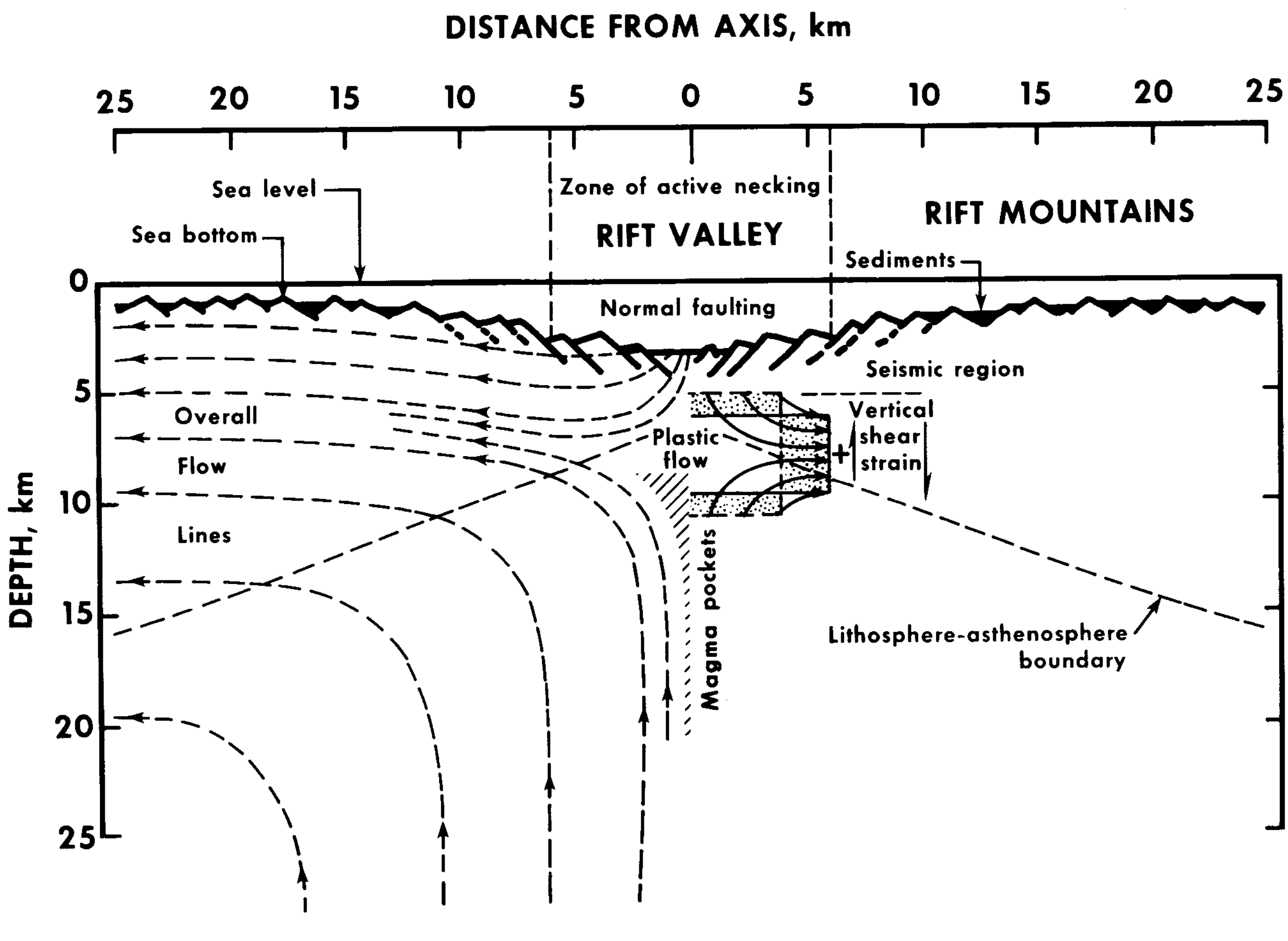
Francis (1974) has proposed a model in which a rift valley is a zone of constant caldera collapse. The inner floor is viewed as the top of a lid floating on a magma chamber and is, at the time of collapse, decoupled from the adjacent rift valley walls. Caldera collapse may occur locally in the inner floor and is observed in Ardoukoba (Needham *et al.*, 1976), but it cannot be a dominant process, since active normal faults are observed at any scale over a broad width (10–15 km) and magma chamber widths are (supposedly) small (< 1–2 km) for slow-spreading ridges (Kusznir and Bott, 1976; Sleep and Wolery, 1975).

Tapponnier and Francheteau (1978) have proposed that because an oceanic rift valley is a steady-state feature, necking must be maintained in a steady-state situation. In a typical extension or squeezing experiment, the sample that is pulled shrinks in the region of necking down to zero thickness or to fracture. In the case of the rift valley, this situation is never reached, since a constant thickness of the lithosphere (4–5 km) is preserved near the axis of maximum necking because of a steady-state temperature gradient. The stretched lithosphere is healed from below through cooling and from above through lava accumulation. Cracks that provide access for the magma to the surface occur in the neck but heal rapidly.

Steady state is achieved only statistically, since the accretion process is a series of discrete successive events (normal faulting, dike intrusion, and lava accumulation; see, for illustrative sketches, Bayer *et al.*, 1973; and Daignieres *et al.*, 1975). The downward-diverging flow lines near the axis of Fig. 6.2.3 illustrate the cumulative effect of combined normal faulting, fissuring and lava piling (Palmason, 1973; Daignieres *et al.*, 1975). Also shown in Fig. 6.2.3 are the flow lines for plastic horizontal extension (Backhofen, 1972) in which vertical thinning balances horizontal stretching. The plastic flow field provides a consistent link between the downward diverging flow in the crust and the upward diverging flow in the mantle.

A consequence of the necking of the lithosphere is a redistribution of mass in the axial region. When the rift inner floor is affected by normal faulting as a result of necking, a small graben is created that results in local lowering of the floor relative to the walls, whereas at a greater depth, plastic deformation takes place. The effect of the plastic deformation is to transfer mass sideways so as to conserve volume (Backhofen, 1972; Fig. 6.2.3, dotted area). Since the density contrast at the bottom of the 5-km layer is negligible, the net result is a mass deficit (water replacing basalt) at the surface in the down-faulted region of the new graben. A net upward force due to buoyancy will affect the layer on top of the asthenosphere in the region immediately below the graben. Since the top layer is pictured as being continuous and with strength, the isostatic balance will not be restored locally but will possess a degree of regionality: the top layer will "bend" upward, and the wavelength of the response will be wider than the graben itself. The topography of nonrifted ridges is not addressed by these models except by that of Lachenbruch (1973).

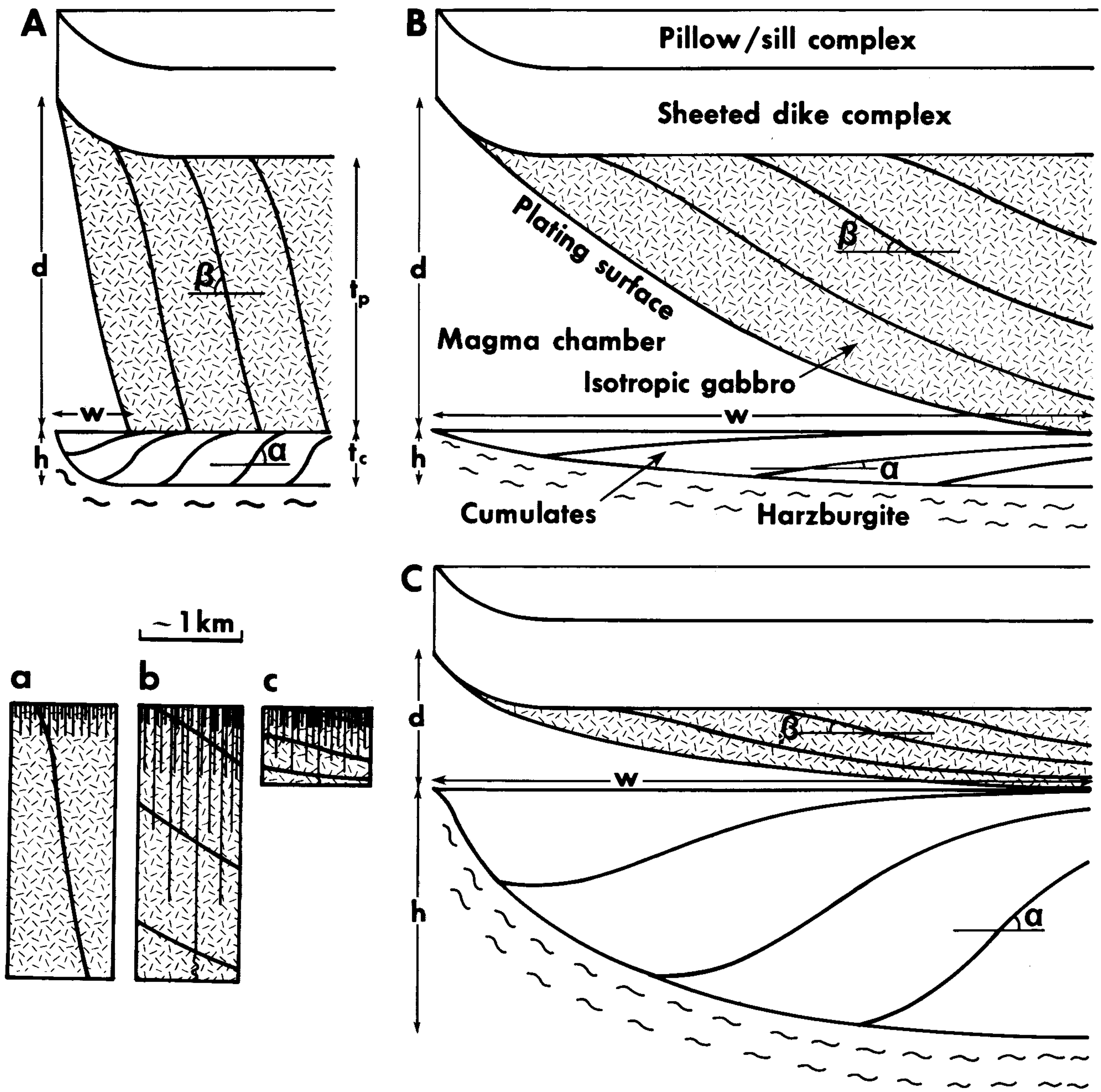
Models that address the lithological assembly of part or all of the oceanic crust and upper mantle include those by Cann (1974), Greenbaum (1972), R. Kidd (1977) and Dewey and Kidd (1977). The treatments of R. Kidd (1977) and Dewey and Kidd (1977) basically agree on the large-scale features of the accumulation of the upper part of the crust (lavas, sills and dikes of basalt and diabase), although which of the various possible detailed geometries of accretion is dominant in this part of the crust cannot be decided from present data. All proposed models depend on the presence of a permanent magma chamber beneath the spreading axis. They differ significantly on the geometry of such a chamber. Dewey and Kidd (1977) also require the lower parts of the cumulate rocks produced from the chamber



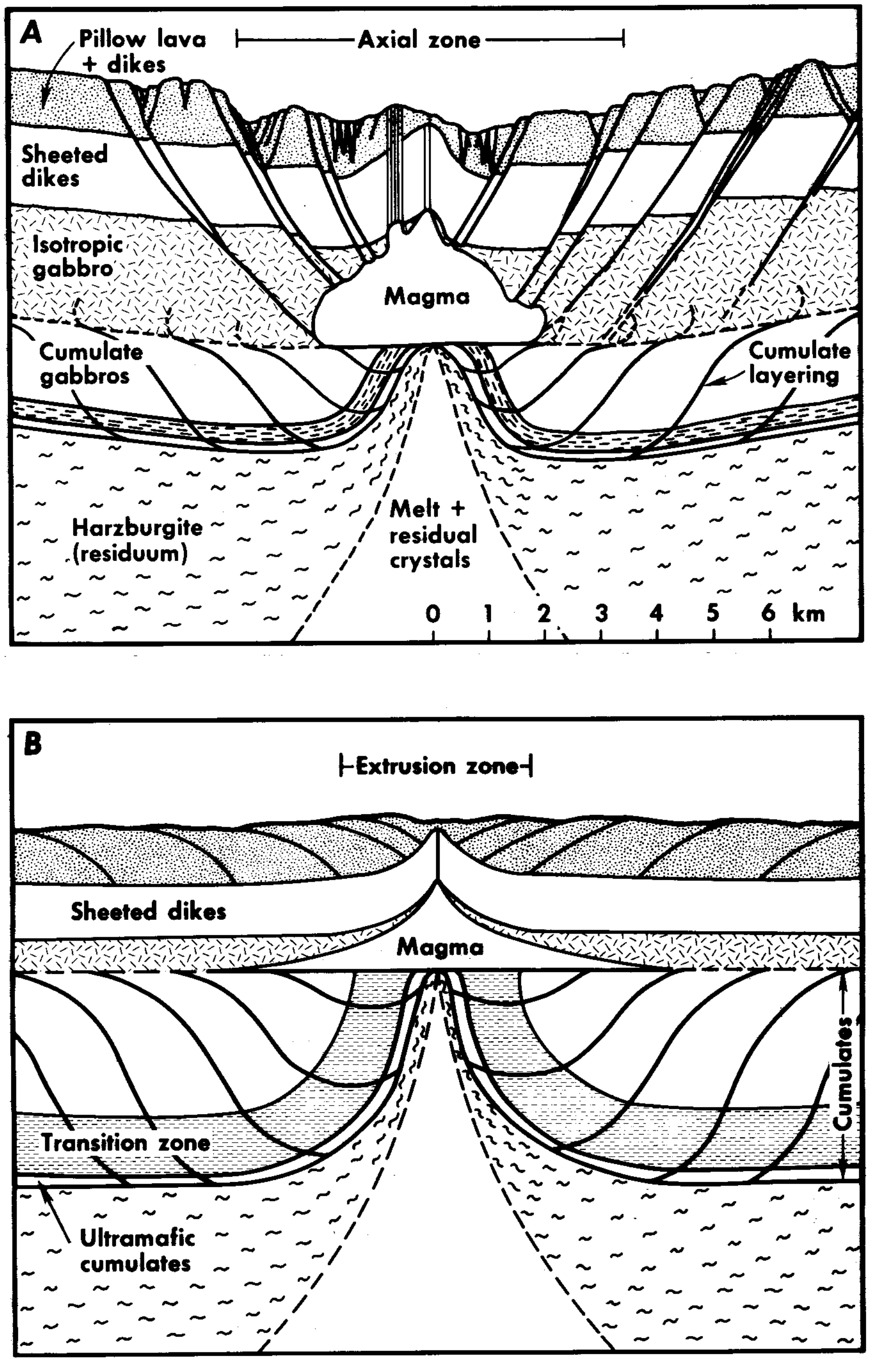
6.2.3 Major features of the mechanical behavior of the lithosphere in the region of plate accretion at plate boundaries with slow spreading rates. Steady-state necking of the lithosphere is suggested to occur. After Tapponnier and Francheteau (1978).

to undergo plastic deformation at high temperature; evidence seen in ophiolite complexes supports the latter contention. The main components of the model suggested by Dewey and Kidd (1977) are summarized in the following paragraphs (see Figs. 6.2.2–6.2.5).

Pillow lavas accumulate progressively, in any one section, from essentially zero thickness at the axis to full thickness 1 or 2 km from the axis. The thickness of 100% sheeted dikes (typically 1 km) is controlled by the depth of hydrothermal circulation at the axial singularity. At least 300 m of gabbro is plated, starting at the axial singularity, onto the roof of sheeted dikes in a complementary, progressive manner to the lavas above. As this gabbro is plated, it must be rapidly cooled by the



6.2.4 Diagrams showing how magma chamber width (w) and depth (d), and the height of the axial welt (h) control the thicknesses of roof-plated gabbro (t_p) and of cumulates (t_c), and the angles (α, β) between small-scale layering and the horizontal. **a**, **b**, and **c** show suggested distribution of dikes in roof-plated gabbro of **A**, **B**, and **C**, respectively. After Dewey and Kidd (1977).



6.2.5 Sketched cross sections of plate accretion model drawn on (**a**) rifted, slow-spreading ridge profile, and (**b**) nonrifted, fast-spreading ridge profile. Profile in **a** from North Atlantic, 36°50′ N (ARCYANA, 1975). Profile in **b** redrawn from Larson and Spiess (1969); East Pacific Rise, 21°N. After Burke and Kidd (1980). Both diagrams at the same scale; no vertical exaggeration.

hydrothermal circulation and, at least near the axial singularity, cut by tensional cracks that are filled by dikes.

The magma chamber is zoned, with dominantly olivine cumulates being deposited near the magma supply axis, grading out to dominantly clinopyroxene-rich cumulates that in turn pass into clinopyroxene-plagioclase cumulates, which continue to the edges of the magma chamber. Magma chamber zonation is strongly suggested by the spatial variation of the composition of erupted magmas (Ballard *et al.*, 1975; Bryan and Moore, 1977). The fact that all well-preserved ophiolite complexes consistently show the vertical sequence of gross layering required by this zonation indicates that the thermally controlled system and the environment producing it are more or less steady-state, and that rapid convective overturn and total mixing are in some way inhibited. If the supply of new magma to the chamber is roughly balanced, on a short time scale (100 yr), by eruptive removal and by the volume created by spreading, then new hot magma should rise directly to the apex of the chamber and should not grossly disturb the somewhat cooler magma nearer the sides. The fact that the zonation exists is a strong argument for a fairly continuous supply of new magma; this is the only reasonable mechanism to keep the central zone in the magma chamber hotter than the sides when the whole roof is being cooled (mainly by hydrothermal circulation). D. Walker points out (pers. comm., 1980) that picritic magma, though hot, may not rise because it is too dense and thus enhances chamber zonation.

The modal compositions and relative volumes of the plated and cumulate rocks show that the phases involved in crystal fractionation are dominantly clinopyroxene and plagioclase; olivine is dominant near the axis but rapidly becomes negligible away from the axis and is, overall, a subordinate phase. Orthopyroxene is always a minor phase; opaque minerals are conspicuously absent from all the rocks, except for chromite in the dunite. Because fresh magma is being supplied fairly continuously to the chamber, fractionation models must allow for some mixing between this and fractionating magma. The magma chamber, above the section in which cumulates are forming, must be filled with magma (liquid + crystals), not with mush (framework of crystals with interstitial liquid). Mush may exist only below the cumulate deposition surface and above or outside the roof gabbro-plating-surface.

A flat-floored magma chamber is required, according to Dewey and Kidd (1977), because of the large thickness of cumulate rocks and of the need to form them within a reasonable distance of the axis. To do this with an essentially steady-state process requires that there be a very narrow axial welt consisting of a mush of residual crystals and melt at its center, and of plated residual harzburgite thickening downward and sideways at its sides. The central part of the welt will be about 70:30 in relative abundances of residuum:melt at the depth where harzburgite plating starts and will progressively change upward to near 100% melt at the magma chamber floor. The geometry requires rapid subsidence of the lower cumulates and implies the presence of a narrow zone of high shear strain between rising mush and subsiding cumulates. The precise shape of the axial welt, defined by the cumulate ultramafic-plated (non-cumulate) harzburgite boundary, is at present unconstrained. The shape chosen gives a maximum amount of room to accommodate the cumulate pile and the zonation required in the magma chamber.

The penetrative deformation, which is seen to affect the harzburgite and zones in the cumulate ultramafics and lower gabbro, is suggested not to be due to horizontal shear by lithosphere plate movement on the asthenosphere, and particularly not to horizontal shear between asthenosphere diverging sideways under the ridge axis more quickly than the lithosphere, as has been implied in several previous treatments. It is evident that material transport in the mantle at shallow depths near the axis under the magma chamber must be dominantly vertical in order to compensate for the gap that would otherwise develop between the two diverging plate edges. Therefore, it is suggested that the penetrative deformation is due mainly to vertical shear stress between the rising mush of residual crystals with basalt melt and the subsiding plated harzburgite and lower cumulates. Some deformation in the cumulates may result from slump or flow processes caused by a rise or fall in the level of the top of the axial welt and to bending stresses in the differentially subsiding cumulate pile. Although the shear stress is vertical, the resulting structures end up nearly horizontal because of the geometry of the subsiding welt and cumulate pile, and nearly parallel with cumulate and other layering because even a small shear strain will rotate these markers into near parallelism with foliation and lineation. Episodic rise or fall in the level of the axial welt requires episodically reduced or accelerated shear strain below; this mechanism may be responsible for zones of more strongly and more weakly deformed harzburgites.

There are some problems that cannot be resolved from data available at present. These are as follows: the relative thicknesses of plated gabbro, cumulate gabbro, and cumulate ultramafics; changes in these as a function of magma chamber width; the origin of large trondhjemite bodies; and correlations between spreading rate, magma chamber width and thicknesses of the various units. The amount and composition of magma supplied for a given increment of spreading on normal ridge segments appear to be fairly constant, because the total thickness of crust in normal oceanic situations is fairly consistent at 6–7 km, and because compositional variations of erupted magmas are small along the spreading ridge and insignificant compared with the variation in composition at any one site.

This model, as illustrated in Fig. 6.2.2, has been deliberately drawn with approximately equal thicknesses of plated and cumulate gabbro. Their combined thickness is shown in this figure in the same proportion to the combined thickness of dikes and pillow lava as oceanic seismic Layer 3 (broadly speaking, gabbro) to oceanic seismic Layer 2 (broadly speaking, dikes and pillows). This has been done because the thicknesses and any differences in thickness of plated and cumulate gabbro in different ophiolite complexes are so poorly known. The model is easily modified (as illustrated in Fig. 6.2.4) to accommodate any necessary changes in the relative thicknesses of plated and cumulate gabbro when these are defined from study of thick, well-preserved ophiolite complexes.

Large-scale thermal models for the cooling of the oceanic lithosphere (Sleep, 1975; Lister, 1977) do not have the resolution necessary to provide the details of crustal accretion even if hydrothermal circulation is included in the analysis. These thermal models do, however, give information on whether it is possible to sustain a steady-state magma chamber for a given spreading rate. The faster the spreading rate, the more likely it is that such a magma chamber exists, but it is not yet clear from the thermal model calculations whether spreading rates typical of rifted ridges (1–2 cm/yr half-rate) allow one to exist. If it turns out that a magma chamber is not a steady-state feature at low spreading rates, then the well-preserved ophiolite complexes are likely to be a skewed sample, representing crust formed only on fast-spreading ridges. However, the lack of extreme fractionation in lavas from slow-spreading ridges is strong evidence for the existence of a steady-state magma chamber along slow-spreading ridges (Shibata *et al.*, 1979).

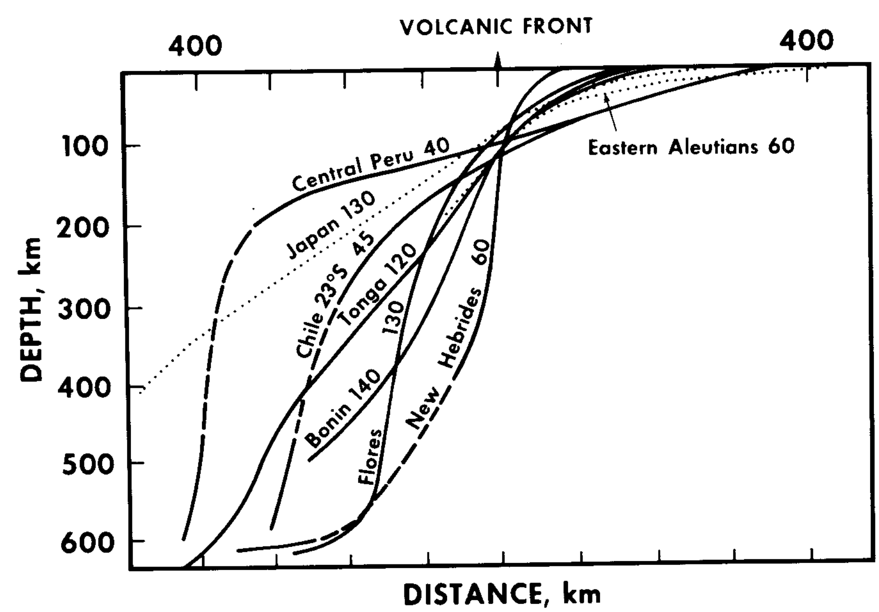
Tectonics of basaltic volcanism at convergent plate margins

The most readily observed active volcanism on Earth is that in Andean and island arcs. Activity is concentrated in a relatively small number of large volcanoes, many of which are subaerial and hence accessible for study, but it is not clear exactly how much of this activity is basaltic. Although andesitic volcanism dominates in arc-related systems, Ewart (1976) has estimated that about 20% of all the igneous rock produced in island arcs is basaltic, and our own estimate is higher, so that the convergent boundary environment apparently rates second (after divergent boundaries) in the volume of basalt currently being produced.

Discussion in this section is divided into considerations of:

(1) the distribution of arc volcanism in relation to subducted lithosphere;
(2) intra-arc tectonic features;
(3) contrasts between island arcs and Andean volcanic arcs;
(4) effects of ridge subduction on arc volcanism;
(5) the origin and evolution of marginal basins in relation to arcs; and
(6) volcanism at arc-continent and continent-continent collisions.

Distribution of arc volcanism in relation to subducted lithosphere. Wadati (1935) recognized that volcanoes in arcs overlie places where the seismic plane, now identified with the descending lithospheric slab, lies at depths greater than 100 km. This relationship has since been observed in many arcs and is now one of the best established associations in plate tectonics. Where the narrow zone of active volcanoes is separated by only a short distance (l) from the trench, the average dip of the seismic zone (defined by the angle θ from the horizontal) is steep and θ is large; conversely, where l is large



6.2.6 Volcanic arcs generally are located 100–150 km above their controlling subducted slabs. Numbers on the eight slab traces indicate the age (m.y.) of ocean floor being subducted at the respective trenches. Based on a compilation by J. F. Dewey.

(up to a maximum of 500 km in the Makran), θ is small. Estimates of l and θ from numerous arcs indicate that the volcanic zones generally occur 100–150 km above the seismic plane (Fig. 6.2.6), and it is therefore inferred that lithosphere descending at these depths is involved in production of arc magmas, although the question of exactly how and where the magmas are produced is as yet unresolved. Although the simple relationship works well, changes in the dip of the subduction zone can also be accommodated (Karig *et al.*, 1976).

A second unresolved question is that of the amount of magma produced along or above the descending slab. More precisely, what is the rate of production of new crustal material at convergent boundaries? The reason why these questions are unresolved is that most of the obvious ways of addressing them are subject to error. Examinations of volcanic rocks erupted in an arc through time have yielded some general proportions and quantities (Sugimura, 1968; Markhinin, 1968), but sampling errors are hard to assess and preservation is very incomplete and is biased against explosive and clastic material. Inferences about eruption rates based upon the apparent volume of the arc are difficult to make because it is not possible to separate that part of the volume which is maintained dynamically by hot, low-density mantle underlying the arc from the part representing newly produced crust. One possible way to obtain a maximum value for magma production at convergent boundaries would be to assume that magma production is related solely to slab dehydration (Anderson *et al.*, 1976), to assume the slab is uniformly (say, 1%) hydrated for the top 3 km of its 100 km thickness, and to calculate how much energy dehydration could release for partial melting of overlying pyrolite. From this calculation, a rate of crustal production (as basalt or basalt equivalent in calc-alkaline material) might be obtained. It seems likely that this value will be a great deal lower than the rate of crustal production at oceanic ridges.

Variation in composition within arcs has been related to a range of tectonic controls (see discussion below). Miyashiro (1972) has suggested that tholeiitic rocks are more abundant where plate convergence is fast and that alkaline rocks are more abundant where convergence is slow. Correlation of variations in alkali contents of volcanic rocks with depth to the underlying slab has been elaborated by Dickinson and his collaborators (Dickinson and Hatherton, 1967; Hatherton and Dickinson, 1968, 1969; Dickinson, 1970; following earlier work by Rittman, 1953; Kuno, 1966; Sugimura, 1968). Although each area should be considered singly because of local complexities, a general increase in alkalies in igneous products with increasing slab depth seems established.

Tectonics within arcs. Arc igneous activity is closely linked with local extensional phenomena. Compressional tectonics (operating in the arc-trench gap between the volcanoes and the site of subduction in the trench) have metamorphic but few igneous phenomena associated with them. Nakamura (1978) has suggested that dike orientation in various arcs can be interpreted as indicating the direction of maximum extensional stress within the arc and that the elliptical shapes of many calderas can be used in a similar way.

Very large calderas are prominent features of Andean arcs, especially in Sumatra and the Peruvian Andes. These structures imply extreme thinness of the lithosphere along the line of volcanoes; the localized lithospheric weakness of this narrow zone is important in the development of marginal basins.

Along-strike tectonic variations in arcs have been recognized by several workers and have been related to igneous activity. A fairly regular 75-km spacing of active volcanoes has been alleged by Marsh (1974) in the Aleutians and Lesser Antilles which he associates with a characteristic thickness of the magma source layer. Segmentation in seismicity and faulting have been linked by Stoiber and Carr (1973) to variations in volcanic behavior, but doubts have been cast on this correlation (e.g., by Isacks and Barazangi, 1977). A number of within-arc variations can be related to corresponding variations in the properties of the underlying slab, such as the subduction of shallow, buoyant ocean floor (Vogt *et al.*, 1976) and the subduction of lithosphere modified by the occurrence of prominent transform fault zones (Ravenne *et al.*, 1977). Both appear to generate within-arc tectonic variations (such as variations in faulting style and elevation).

Contrasts between island and Andean arcs. Island arcs are constructed on oceanic lithosphere and Andean arcs on continental lithosphere. Where marginal basins form, Andean arcs may become sandwiched by oceanic lithosphere and may evolve into island arcs. The character of the continental lithosphere on which Andean arcs form is itself variable. In many places (e.g., the eastern Aleutians) presently active arcs lie on top of Mesozoic arc material accreted to the continent within the last 100 m.y., but in other places (e.g., the Andes of Peru) arcs lie on rocks that have been continental for a large fraction of geological time. Thus, highly evolved continental crust may lie beneath such an area. There are all gradations of arc foundations, from mature, thick, continental lithosphere to oceanic lithosphere. An arc that has been the site of convergence for tens of millions of years apparently develops a substrate similar in age, thickness and related properties to that of a relatively young addition to a continent.

The variety in arc features becomes more understandable when consideration is given to the range of complexity implied by the variety of processes that lead to arc formation. Some features and changes in tectonic style can be satisfactorily related to changes in such properties as the dip of the subducting slab (e.g., Isacks and Barazangi, 1977) and to the age of the subducted lithosphere, but there is still no clear explanation of many prominent arc features. For example, evidence that the Altiplano of Peru was uplifted within the last 6 m.y. by 3 km or more is convincing (Audebaud *et al.*, 1973), but the mechanism by which it was uplifted is not clear; nor do we know yet why parts of the Andean arc are volcanically active today while others are not. Papers presented at the first Ewing symposium (Pitman and Talwani, 1977) give a good idea of general strengths and weaknesses of our understanding of arc activity. Against this framework, it is not surprising that our appreciation of variations in basaltic and related igneous activity in arcs is still problematical.

The most prominent contrast in igneous activity between island and Andean arcs is that silicic (~ 70% SiO_2) volcanism, often in the form of large-volume ignimbrites with high contents of sodium and potassium, is concentrated in—if not confined to—arcs underlain by continental lithosphere (Andes, New Zealand, Japan and Java-Sumatra). Some involvement of the underlying continental crust in the production of these volcanics appears very likely.

The effects of ridge subduction on arc volcanism. DeLong *et al.* (1978) and DeLong and Fox (1977) pointed out that ridge subduction causes a temporary decrease or cessation of arc magmatism while inducing an episode of low-grade regional metamorphism and wholesale elevation of the arc. The effects of ridge subduction have been observed in the Late Eocene to Miocene geology of the Aleutian arc, where the Kula ridge was subducted about 30–35 m.y. ago (DeLong *et al.*, 1978). Kidd *et al.* (1977) reported on a similar event from the medial Ordovician of Central Newfoundland. Theoretical analyses showing why subduction of a hot ridge can be expected to lead to a temporary shut-off in arc volcanism have been presented by DeLong *et al.* (1978). Such topics as the time of initiation of volcanism relative to the beginning of subduction; correlations between rates of extension and rates of subduction; migration of the volcanic arc with time; and changes in chemical characteristics of volcanics with time are perhaps best addressed in the context of a full kinematic analysis of the motions in the arc system (Dewey, 1980).

Origin and evolution of marginal basins. **Marginal basins** are defined as areas with oceanic water depths and crustal structure lying between a continent and a nearby island arc. Karig (1971a) first explicitly recognized that some of these basins are at anomalously shallow depths for their predicted or known ages, that they have steep, faulted margins and contain little sediment. Some high heat flow values have been obtained in marginal basins. Karig drew the conclusion that active or recently active seafloor spreading was responsible for generating these particular marginal basins. Packham and Falvey (1971) came independently to the same conclusion. It is important to distinguish this environment of seafloor spreading from that of the main ocean ridges, because in marginal basins it is a consequence of the existence of the island arc and the underlying subduction zone. The spreading is not a primary part of the global plate-boundary system. McKenzie and Weiss (1975) have suggested that main ocean basins were generated by propagation of rifts from marginal basin sites. This is clearly not applicable to, for example, the early Jurassic opening of the central North Atlantic, nor to the early Tertiary opening of the Labrador Sea and Baffin Bay; thus, it is doubtful that this proposal has any general validity.

Until recently, studies of marginal basins thought to contain active or recently active spreading centers did not report well-defined symmetrical sets of magnetic anomalies such as those that characterize the major oceanic spreading ridges. Also, the topography in these basins is much rougher than that near normal oceanic spreading ridges that have rates of spreading similar to those inferred for the basins. Except for the Andaman Sea (Rodolfo, 1969; Curray and Moore, 1974), no well-defined rifts such as those that characterize most slow-spreading ridge axes have been found. Since the supposed spreading axis or axes cannot be located accurately and because the topographic profiles are very "rough," it is hard to demonstrate the existence of a systematic subsidence pattern for marginal basin crust like that so well-established in the major oceans. Recently, however, fragmentary magnetic anomaly patterns due to seafloor spreading have been detected in several active or recently active marginal basins (Barker, 1972; Weissel, 1977; Watts and Weissel, 1975; Kobayashi and Isezaki, 1976). These patterns seem to indicate that the spreading process in marginal basins is prone to jumping of the ridge axis, particularly in a direction towards the island arc, so that extensive, simple, symmetrical patterns of anomalies should not be expected. Although the rate of spreading has not been determined very precisely in most active marginal basins, Weissel (1977) has given data indicating that the

Lau basin has opened at an average rate of about 5 cm/yr (total rate) near the north end of the basin. Kroenke and Scott (1978) have given data that yield average total opening rates of about 5 cm/yr for the Parece Vela basin and 2 cm/yr for the Mariana Trough. Although there is a crudely defined axial high in the Mariana Trough (Karig, 1971b; Anderson, 1975), except very locally, the well-defined axial rift that might be expected at the inferred spreading rate does not exist. Anderson (1975) has shown that heat flow measurements support the idea of active spreading and an active hydrothermal circulation system on the axial high of the Mariana Trough.

One possible explanation for the rough topography and ill-defined nature of the spreading magnetic anomalies in active marginal basins was given by Kidd (1977), who suggested that widespread tholeiitic volcanism was likely to occur in marginal basins away from the places where crust was accreted. Klein *et al.* (1978) have subsequently found evidence for such activity in the Shikoku basin.

A limited amount of data are available on the geochemistry of basalts from active marginal basins, and more will shortly be forthcoming in studies following Legs 58-60 of the Deep Sea Drilling Project in the Shikoku Basin and Phillipine Sea. Dredged samples from the Marianas Trough have been described by Hart *et al.* (1972), and Hawkins (1976, 1977) and Gill (1976) have documented the chemistry of rocks from the Lau-Havre Trough. Reay *et al.* (1974) have described the rocks from an active volcano (Niuafo'ou) in the northern end of the Lau Basin. Basalts from the actively spreading marginal basin in the Scotia Sea have been described by Tarney *et al.* (1977). All of these investigators have commented on the close similarity of the young basalts in these active marginal basins to the tholeiitic basalts of the better-known oceanic spreading ridges. However, it should be noted that none of these studies has been made with the idea that there may be widespread "off-axis" or subsequent volcanism in marginal basins, which might conceal much of the crust formed at the axis or axes of accretion. It has been suggested (Kidd, 1977) that basaltic and perhaps ultramafic komatiites might be found forming part of the crust in some marginal basins, particularly narrow ones such as are well-developed in the New Hebrides arc (Karig and Mammerickx, 1972). These suggestions follow from the occurrence of basaltic komatiites (Kidd, 1977) and perhaps ultramafic komatiites (Upadhyay, 1978) in sequences of the northwestern Newfoundland Appalachians, which have been interpreted as oceanic crustal floor and as sedimentary and volcanic fill of small marginal basins of Ordovician age. Detailed geochemical information has been obtained from a fossil Mesozoic marginal basin in the southern Andes (Saunders *et al.*, 1979). Basalts there are said to closely resemble oceanic spreading ridge basalts. A relationship of particular importance for students of marginal basin volcanism is the passage of the Lau-Havre Trough southward onto land in the Taupo volcanic rift zone of North Island, New Zealand. This relation, first clearly pointed out by Karig (1970), shows the largest along-strike variations in presently active marginal basin volcanism, with abundant dominantly ignimbrite volcanism of the Taupo zone passing laterally to the wholly basaltic spreading in the Lau basin. It seems likely that the variation in amount of extension, increasing northward along the zone, is connected with the northward increasing subduction rate. At present, it is not clear if the silicic magmas of the Taupo zone are dominantly of crustal or of mantle origin, and hence it is unclear to what extent the continental-type crust of New Zealand modifies the volcanic expression of the marginal basin.

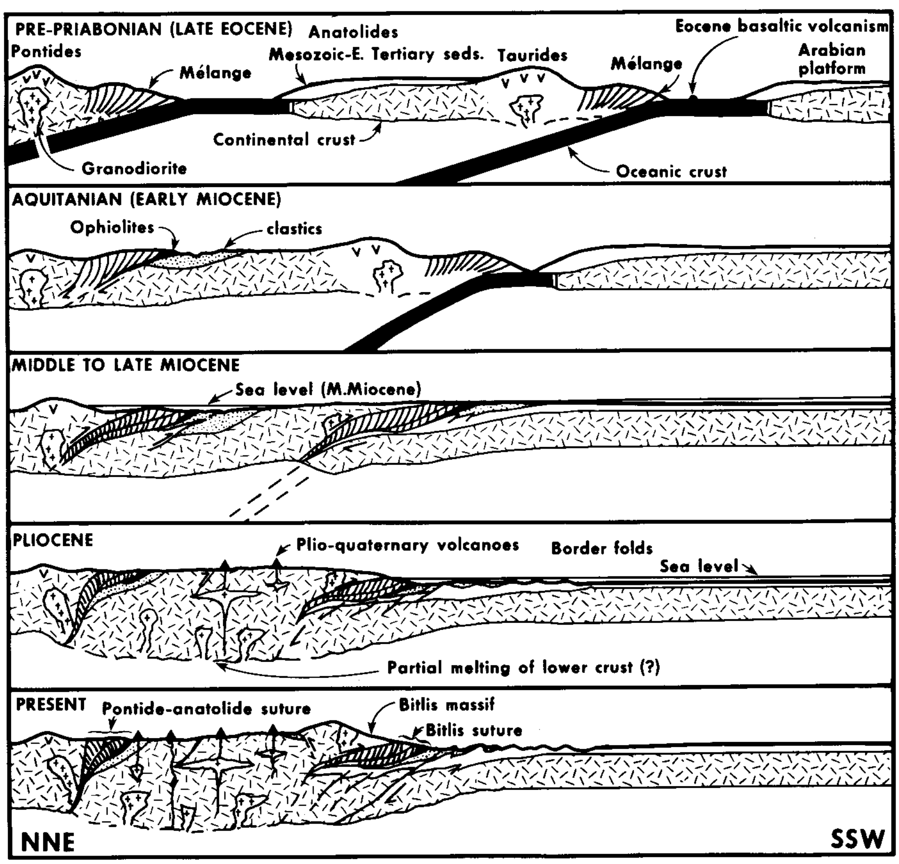
It is important to note that marginal basins formed by the entrapment of oceanic lithosphere (generated in a major ocean by part of the global plate boundary network) between a new island arc and a continent are not discussed in this section. Examples of these "trapped" marginal basins, which often show better-developed magnetic anomalies at high angles to the nearest piece of bounding island arc, are the Bering Sea (Cooper *et al.*, 1977), possibly the West Philippine Basin (Karig *et al.*, 1973) and, perhaps, the South Fiji Basin (Watts *et al.*, 1977).

Volcanism at arc-continent and continent-continent collisions. Active arc-continent collisions are occurring at Timor, eastern New Guinea and Taiwan but are poorly known. Neogene collision of an arc with Australia in New Guinea has been followed by development of subduction in a different direction. This may be a general consequence of arc-continent collision. Continent-continent convergence, however, is persisting after collision from the eastern syntaxis of the Himalaya to Turkey (with the exception of the Makran gap). A recent comparative discussion of the tectonics of Turkey, Iran and Tibet by Şengör and Kidd (1979) illustrates the volcanic complexity that can occur in this environment.

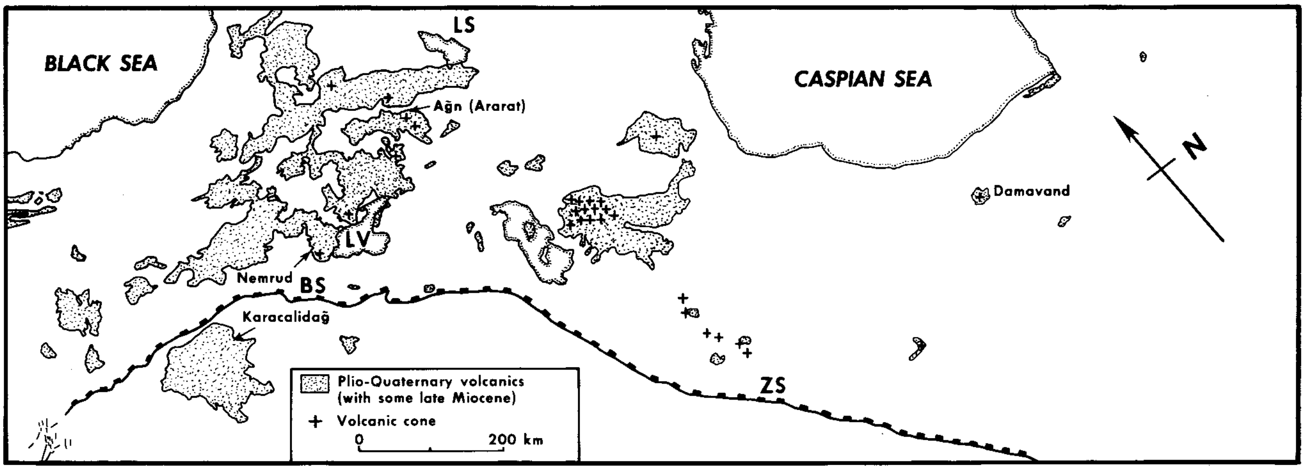
The Turkish-Iranian Plateau is a high region with an average elevation of about 1.5 km. During the late Miocene, the last piece of oceanic lithosphere between the Eurasian and Arabian continents was eliminated at the Bitlis/Zagros suture zone (Fig. 6.2.7). Continued convergence across the collision site resulted in the shortening of the plateau across strike by thickening and along strike by sideways motion. Predominantly calc-alkaline volcanism is present on the highest portions of the area, despite the absence of a descending slab of lithosphere. Surface geology and volcanism of the Turkish-Iranian Plateau greatly resemble those of the Tibetan Plateau.

The post-Miocene tectonic picture of the Turkish-Iranian Plateau is one of active shortening across the plateau. In Turkey a considerable portion of this convergence is taken up by wedging out the Anatolian Plate along the North and East Anatolian transform faults with throws of 85 ± 5 km and 22 ± 5 km, respectively (Seymen, 1975; Arpat and Şaroglu, 1972; Seymen and Aydin, 1972). Active shortening has resulted in some strike-slip faulting to the east of where these faults meet (Toksöz *et al.*, 1977), but thickening is predominant on numerous thrusts. In Iran, thrusts and folds are abundant and probably greatly predominate over sideways motion (see summary in Şengör and Kidd, 1979). This may be because (1) Iran has less space to "escape" into than does Turkey and (2) its lower elevation requires less work for the present uplift. Negative Bouguer anomalies (−150 mgal) over a large part of the Turkish-Iranian Plateau (Özelçi, 1973a, b) are most easily, but not uniquely, interpreted in terms of crustal thickening, especially in the light of the geological data.

The Turkish-Iranian Plateau, especially in its western and central most elevated sections, is also the locus of intense late Tertiary-Quaternary volcanism (Fig. 6.2.8) and contains volcanoes that have erupted during historic times (e.g., Mt. Nemrud eruption, 1441 A.D.; Erinç, 1953). Although parts of the region were characterized by Cretaceous to Miocene calc-alkaline volcanism prior to the Pliocene to Recent volcanic phase (Ketin, 1961; Altinli, 1966), this was probably related to a subduction zone consuming Bitlis-Zagros ocean floor and dipping north-northeast under the future Turkish-Iranian Plateau as shown by the progressive increase in K_2O/SiO_2 ratios in the associated igneous rocks from south to north across the plateau (Adamia *et al.*, 1977). This phase of volcanic activity came to an end in Turkey about 20 m.y. ago (Innocenti *et al.*, 1976) and in Iran sometime during the late Miocene (Jung *et al.*, 1976). Volcanic activity recommenced during the Pliocene and is still active (Ketin, 1961; Gansser, 1966). There is no evidence of a descending lithospheric slab beneath the plateau that can be connected with the active volcanism; the age of collision (approximately



6.2.7 Schematic sections illustrating the evolution of the Turkish-Iranian plateau where active basaltic volcanism is related to continental collision. After Şengör and Kidd (1979).



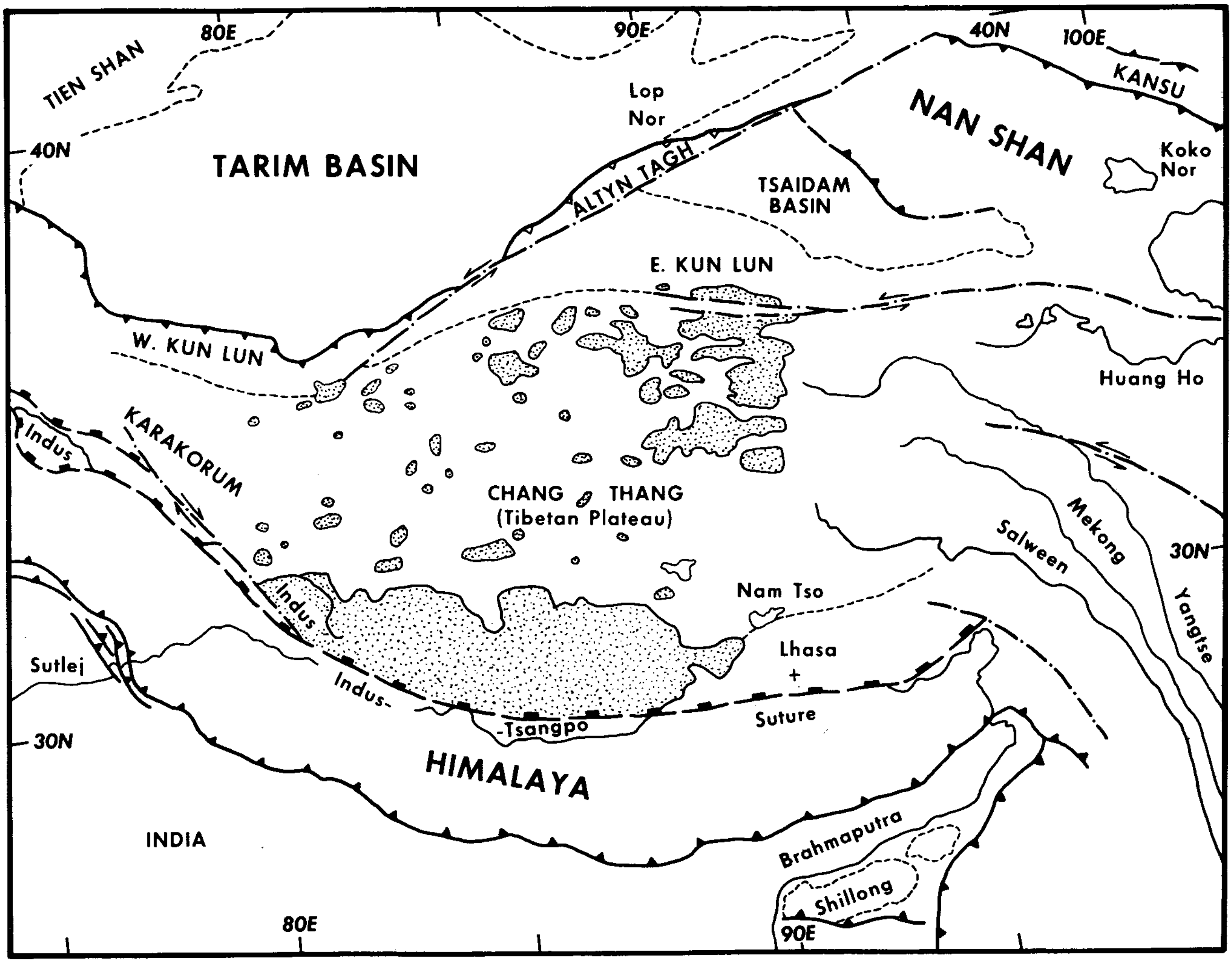
6.2.8 Distribution of Pliocene to Quaternary volcanic rocks on the Turkish-Iranian plateau. This volcanic activity is of great compositional range and is associated with collision between the Arabian continental mass (at the bottom of the map) and the complex arc and micro-continental terrains of Turkey and Iran across the Bitlis Suture (BS) and the Zagros Suture (ZS; LV is Lake Van; LS is Lake Sevan). Modified from Şengör and Kidd (1979).

15 m.y. ago) and the average convergence rate over this time period (about 4.5 cm/yr, McKenzie, 1972) indicate that the slab must long be past the 100–150 km depth where the majority of the calc-alkaline melts are generated. Pliocene to Recent volcanism is very extensive in the highest parts of the plateau (Fig. 6.2.8) and includes both calc-alkaline and alkaline associations, although the former greatly predominate over the latter. The calc-alkaline association is represented by andesites, dacites and rhyolites with some ignimbrites, whereas basalts and very limited phonolites and trachytes represent the alkaline association. Some volcanoes (e.g., Nemrud; Özpeker, 1973) appear to have erupted both alkaline and calc-alkaline rocks. Lambert *et al.* (1974) have reported a $^{87}Sr/^{86}Sr$ ratio of 0.7050 ± 0.0005 from the calc-alkaline lavas of Mt. Agri. If the Devonian and Permian rocks that outcrop very near the volcano represent the total basement beneath it, then this ratio precludes the continental crust as the source material for the melt. But in a terrane as complex as eastern Turkey, which contains a huge amount of accretionary mélange material, this may not be the case. The observations that the Plio-Quaternary volcanics are almost entirely confined to the highest parts of the plateau and that there is a zone of seismic attenuation beneath the plateau (Toksöz and Bird, 1977) support the view that a great portion of the calc-alkaline volcanics here may be products of the partial melting of lower levels of the thickened continental crust.

Over large areas of the Tibetan Plateau, as far as 92° E, the deformed pre-Mesozoic rocks are unconformably overlain by a sequence of Jurassic and Cretaceous sedimentary rocks. Clastic rocks, mainly red sandstones at the base, pass up into rudist-bearing limestones. In the northern part of the plateau the rocks are mainly red sandstones and locally contain gypsum. These Mesozoic sediments are found over the whole plateau south to near the Indus Suture, including near Lhasa, where they are strongly folded (Hayden, 1907). Dips recorded by Hedin (Backstrom and Johannsen, 1907; Hennig, 1915) and analysis of Landsat images show that the Mesozoic sequence is buckle-folded, in places very strongly so, over the whole width of the plateau, with east-west trending axial surfaces. Folds of this type are usually accompanied by thrust faults (e.g., in the Jura Mountains), although it is difficult to identify these positively from the imagery. In the Tsaidam Basin northeast of Tibet (Fig. 6.2.9), the youngest playa sediments appear to be folding now and are overthrust by the Permian rocks on the northeast border of the basin. The axis of maximum shortening, as judged from the strike of the fold axial planes, is northeast as opposed to the north-south shortening axis given by the east-west axial planes of Tibet. This may be due to a recent reorganization of the active deformation as it spreads away from the relatively stable Tarim Block. In terms of style of folding, overall morphology and tectonic setting, the Tsaidam Basin is very similar to the Dasht-i Kavir Depression of Iran.

Young volcanic rocks of andesitic and more silicic calc-alkaline compositions are widely developed over the Tibetan Plateau as far as 92° E (Backstrom and Johanssen, 1907; Hennig, 1915; Norin, 1946; Burke *et al.*, 1974; Kidd, 1975). Those in the northern and central parts of the plateau clearly postdate the folding of the Mesozoic strata in many places. They are known to consist of both alkaline and calc-alkaline potassic lavas (Deng, 1978). In the southern part of the plateau, a 200-km-wide belt of volcanics adjoins the northern side of the Indus Suture, and stretches from the Indus in the west to near Lhasa in the east. It contains a few volcanic features which are obviously young, judging from the LANDSAT imagery, but the youth of most of these abundant volcanics is not morphologically obvious. Specimens collected by Hedin from this terrane are mostly ignimbrites and subordinate related igneous rocks. While this belt could be the remains of an Andean-type magmatic arc, the reports of abundant hot and boiling springs in this terrane indicate the widespread presence of magma at shallow depth. For this reason, and because draping relations to folded Cretaceous sediments can be seen on the Landsat images in the northern part of this terrane, it has been suggested that the bulk of these volcanics are young (Neogene and younger), although a small portion could be of early Tertiary or late Cretaceous age (Burke *et al.*, 1974). The uplift of Tibet, although poorly dated, clearly predated the ongoing uplift of the Himalayas; this is shown by the antecedent Indus and Brahmaputra rivers.

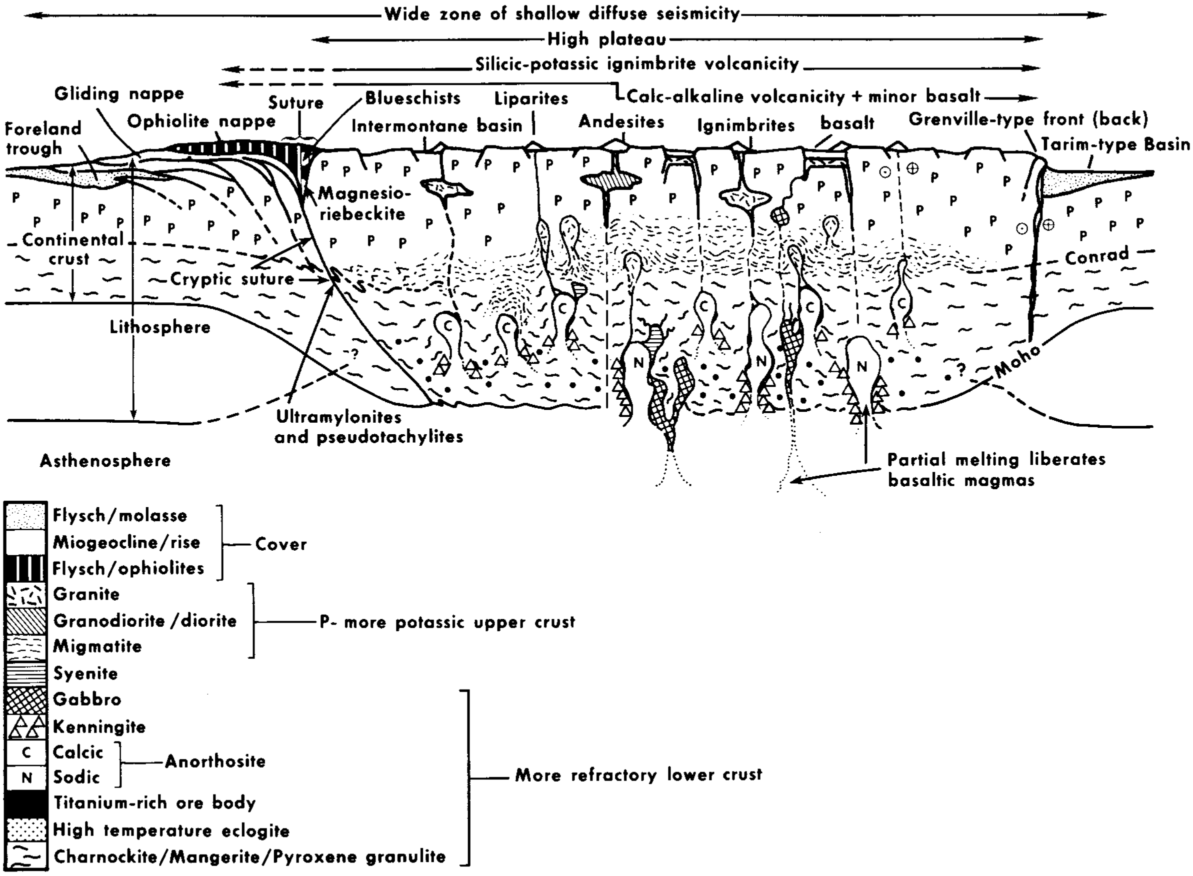
The great resemblance between the Turkish-Iranian Plateau and Tibet, with respect to overall morphology and tectonics, was first emphasized by Von Zahn



6.2.9 Stippled areas represent Neocene and more recent (including active) volcanism of great compositional range over nearly 10^6 km^2 associated with the collision of India with Asia. Lines with black triangles are active thrust faults; lines with open triangles are inactive thrust faults; lines with black rectangles are suture zones; and dash-dot lines are strike-slip faults. Modified from Şengör and Kidd (1979).

(1906). The foregoing descriptions show that the geological resemblances between the Tibetan and the Turkish-Iranian Plateau are readily apparent. Both plateau areas adjoin a suture where continental apposition has occurred and collision is in progress. The earlier onset of the collision, perhaps 30–40 m.y. ago (Dewey and Burke, 1973; Molnar and Tapponnier, 1975), is perhaps one reason why Tibet is higher and more extensive than the Turkish-Iranian Plateau. In both areas the plateaus were uplifted before the suture zone formed. We take this and the available seismic evidence to indicate that large-scale continental underthrusting, as suggested by Argand (1924), is not the cause of uplift. Shortening and resultant thickening, as proposed by Dewey and Burke (1973; Fig. 6.2.10) and favored by Le Fort (1975) for the origin of Tibet, appear to be the model most consistent with the currently available data from the Turkish-Iranian Plateau, as well as those from Tibet. Therefore, the lower elevation of the Turkish-Iranian Plateau, especially of the Iranian segment, may be also due to the existence of cratonic nuclei within the Central Iranian Plateau (Stöcklin, 1974) that resist deformation by shortening better than the relatively weaker accretionary prism material that appears to make up a large portion of the Tibetan basement.

Both plateau areas exhibit folding of covering sedimentary rocks in at least part of their areas. In the Turkish-Iranian Plateau, this folding, accompanied by



6.2.10 Basaltic and other volcanism produced at continental collision. Collision of two continents commonly results in the thickening of one, producing a Tibet-like plateau. This thickening is produced by horizontal shortening including thrusting and folding at shallow levels not shown on figure. The thickened continental crust is raised in temperature mainly by being depressed into hotter regions. The complex igneous and metamorphic results that follow are depicted in this figure. Partial melting of the continental crust yields highly-potassic volcanic rocks and shallow intrusives, leaving refractory residue at depth. Basaltic volcanism develops where the mantle below the thickened continent (>70 km depths) becomes sufficiently hot. Note that the lithosphere-asthenosphere boundary is poorly defined under the elevated plateau and may be no more than a few kilometers down. This style of modification represents an important stage in continental evolution that has happened widely but not universally. Areas that have escaped collision preserve Andean and island arc characters. Modified from Dewey and Burke (1973).

extensive thrusting, started nearly synchronously with the collision. The time of the folding in Tibet is less well-constrained, but the huge area affected by folding, particularly across strike, is remarkable and suggests that it is unlikely to have happened before the collision, considering the active folding in the Tsaidam Basin. Thrusting in Tibet does not seem to be as widespread as it is in the Turkish-Iranian Plateau, but we believe this to be an artifact of recognizing thrusts on Landsat imagery and not the result of the actual absence of the process. Both plateaus show minor normal faulting at high angles and strike-slip faulting at low angles to the suture.

Volcanism on the Tibetan Plateau greatly resembles that found in eastern Turkey and western Iran. It is similar in its composition, its wide extent, its occurrence on high ground, and, at least for a large proportion of the Tibetan volcanics, in its post-collisional age. Alkaline volcanics are present on both plateaus, although these are subordinate in amount to the very potassic calc-alkaline volcanics. From the point of view of basaltic volcanism, the remarkable thing about geologically young, collisional, orogenic volcanism is the great rarity of basalts relative to more silicic rocks. The only other comparable volcanic environment is the altiplano of the Andes, where similar volcanism occurs on an elevated plateau, but without accompanying continental collision.

Tectonics of basaltic volcanism at transform plate margins

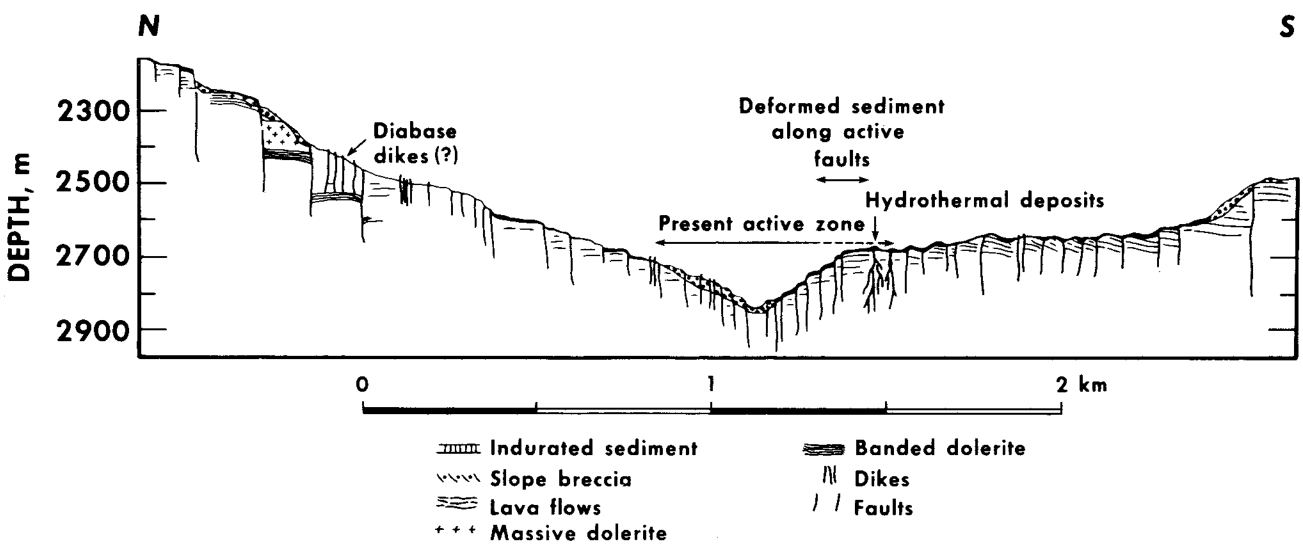
Summary of transform fault tectonics. **Transform faults** were recognized and defined by Wilson (1965a) as plate boundaries across which lithosphere is undergoing relative displacement but is neither being created by seafloor spreading nor destroyed by subduction. This hypothesis was quickly and elegantly confirmed by the studies of earthquake first motions made by Sykes (1967). By implication, because extension and compression are not primary features of transform faults, Wilson's hypothesis suggested that volcanism would not be found to be a major component of their tectonics. Secondary extension or compression may occur across irregularities in the trends of transform faults. These local minor departures from pure strike-slip tectonics, when they lead to extension, may give rise to volcanism in some places along transform faults. Reasons for such secondary tectonics on transform faults have been discussed by Menard and Atwater (1968, 1969), Van Andel *et al.* (1971) and Dewey (1975). Another minor departure from pure strike-slip tectonics on oceanic transforms is the production of large, elongate ridges flanking one or, in places, both margins of the transform trough (Bonatti, 1971, 1973; Thompson and Melson, 1972; Bonatti *et al.*, 1979). These consist mostly of peridotite and serpentinized derivatives which may originally be intrusive, but which almost certainly have been uplifted by tectonic means, probably aided by, or perhaps largely due to, density reduction through serpentinization. Although some authors have proposed that sections of normal oceanic crust and upper mantle may be exposed on the slopes of these fault-bounded ridges and on the walls of transform and fracture zone troughs, Francheteau *et al.* (1976) convincingly demonstrated that, in general, this is unlikely to be true. A complex distribution of lithologies along transform faults and their fracture zone extensions is suggested by data from the Oceanographer Fracture Zone (Fox *et al.*, 1976). The complexity presumably results from the protracted history of tectonic slicing and differential horizontal and vertical displacements that rocks may suffer within the transform domain. Furthermore, it has been suggested (Fox, 1978; Gallo and Fox, 1979) that large-scale thermal effects are responsible for significant changes in oceanic crust near transform faults, due to the abutment of older, colder lithosphere against spreading ridge ends. The most prominent change suggested—a general marked thinning of both of the basaltic layers (and hence of the overall crust)—has been confirmed in at least one study (Detrick and Purdy, 1978). Therefore, even if essentially intact sections of oceanic crust are exposed locally in transform or fracture zone walls, the thickness and perhaps the lithologies and their sequence will not be representative of oceanic crust away from transform faults and fracture zones.

Volcanism on oceanic transform fault/fracture zones. Apart from volcanism at the ends of spreading ridges where they join transform faults, the evidence for volcanism along oceanic transform faults and their fracture zone extensions is not abundant. The young alkalic basalts and gabbros dredged along the St. Paul and Romanche fracture zones (Melson *et al.*, 1967; Honnorez and Bonatti, 1970; Bonatti *et al.*, 1971) are the only rocks now known that are likely to have been erupted on transform/fracture zones away from the places where spreading ridge axes abut such zones. Even in these two cases, the tectonic details of their occurrence and the local abundance of the alkaline volcanics (or lack of it) are poorly known. It is also hard to prove that these volcanics are not related to areas of "hot spot" volcanism, rather than to general transform/fracture zone tectonics. This region could be a hot spot with relatively little associated volcanism, like the Jos Plateau and Air regions of the African plate. The latter are clearly identifiable since they are not submarine, lie within a plate, and are obviously associated with a structurally and topographically defined uplift. Perhaps the fact that there is an uplift at St. Paul's Rocks favors the idea that these volcanics are hot-spot-related rather than related to secondary extension across the transform/fracture zone, since the presence of an anomalously uplifted block on a transform fault might be taken to indicate that the zone is under compression, a condition not favorable for volcanism.

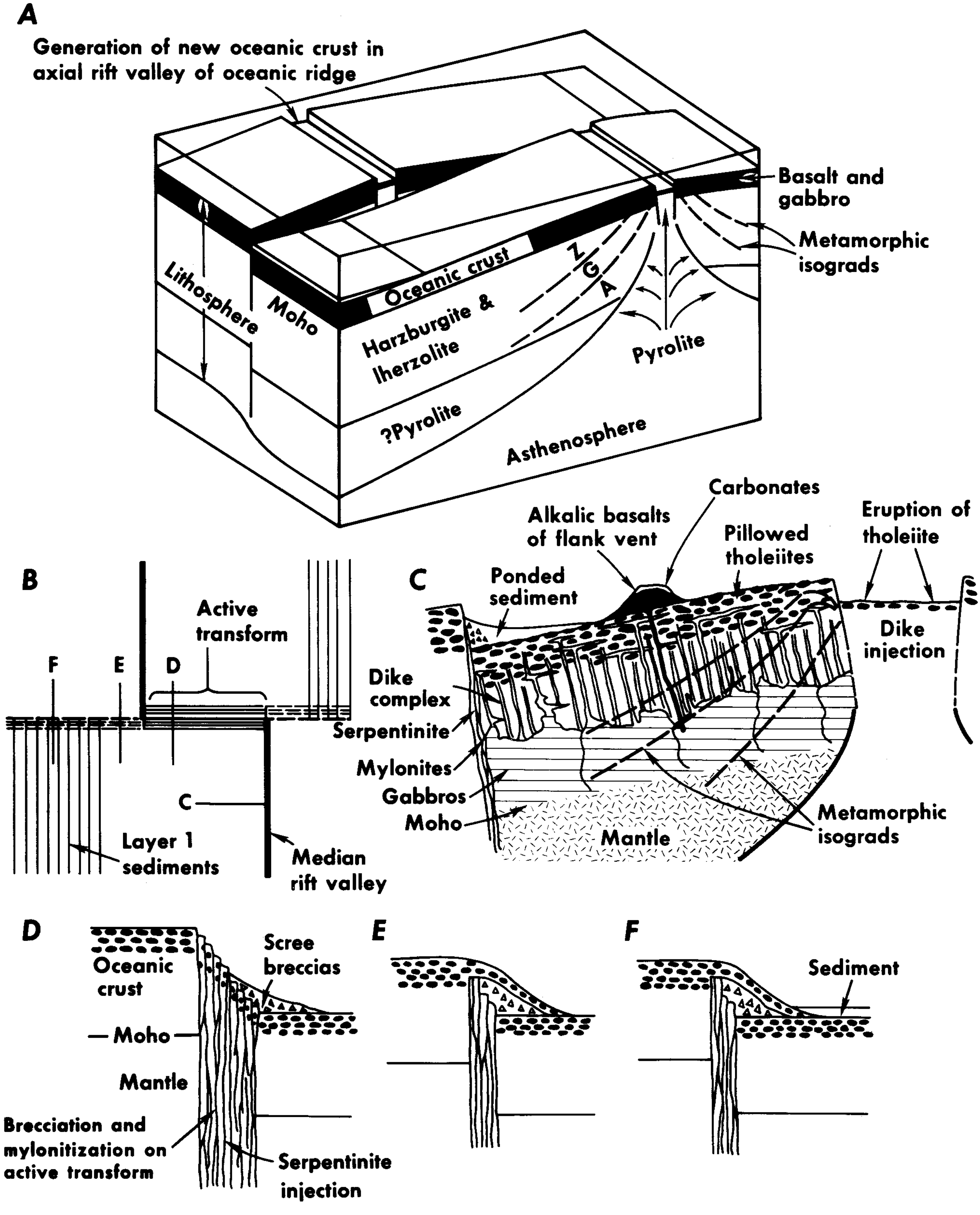
ARCYANA (1975) and Choukroune *et al.* (1978) observed a few wall-like features trending parallel to the transform in the outer, less tectonically active part of the transform valley at the northern end of the FAMOUS area (Fig. 6.2.11). These were tentatively identified as dikes that had just reached the surface but failed to produce flows. If true, this indicates that some very minor magmatism occurs at some places along the transform zone. However, as these dikes (?) have thick manganese coatings, they could have formed close to the ridge axis. It also seems possible that they are lithified breccia zones rather than dikes, so their significance is uncertain.

Other evidence for young volcanism on oceanic transform fracture zones is indirect, consisting of seamount massifs that occur on the Oceanographer and Kane Fracture Zones. In the case of the Oceanographer Fracture Zone, the seamounts are located at the place where there is a significant change in the orientation of the fracture zone (P. J. Fox, pers. comm., 1978). This change is such that there is likely to have been an extensional zone developed during the change in spreading direction recorded by the different orientations of the fracture zone. Such an extensional zone, i.e., one that results in volcanism but not in the establishment of a new spreading ridge segment, has been termed a "leaky" transform; however, there are no well-documented examples of currently active "leaky" transform faults in the ocean, and there are few examples on land. The term should probably not be used for places where a short, spreading ridge segment joins two long transform faults, such as the Cayman Trough. It seems that even small extensional zones such as this develop a moderately coherent spreading center geometry, although there is probably some critical size range below which the intrusion and extrusion of basalt are less regular; such pull-apart zones without coherent spreading centers may legitimately be called "leaky transforms." It is unlikely that these leaky transform zones will persist for very long. They will disappear either because a further slight change in relative plate motion removes the component of extension across the transform, or because new ridge-transform segments develop which are more concordant with the new spreading direction. The ephemeral character and probable small scale of "leaky" transforms may be the reasons why current examples are hard to find in the oceans.

Most basalts dredged from oceanic ridge-ridge transform faults and fracture zones most probably erupted near one or the other of the two places where spreading ridge segments abut the fracture zone-transform fault intersections. Such basalts are more properly regarded as products of the spreading ridge



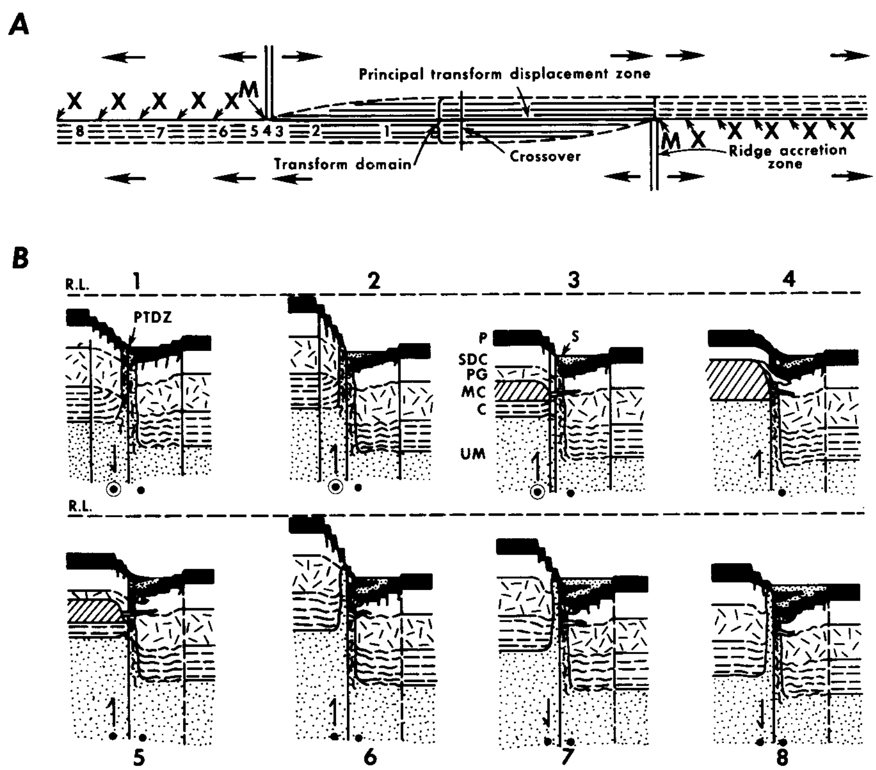
6.2.11 Diabase dikes (?) observed in the FAMOUS area are associated with a transform fault. See text for details. Slightly modified from ARCYANA (1975).



6.2.12 Block diagram (**a**), and map (**b**) showing ridge-transform-ridge intersection. **c**. Detail of accreted oceanic crust. **d**–**f**. Proposed structures and stratigraphic relationships developed along a transform fault and fracture zone; position of sections given in **b**. After Dewey and Bird (1971).

segments; however, interesting and complex relationships between basalts and the products of transform tectonics are likely to be developed in this system (Karson and Dewey, 1978). Two relationships are likely to be common, the first being that predicted by Dewey and Bird (1971; Fig. 6.2.12): this is the production of scree breccias on fault scarps in the transform zone which then become overlain by lavas as the transform-affected crust passes the end of the spreading ridge. Pelagic sediments might also be incorporated in similar sequences, and interbedded pelagic sediment and volcanics were observed at a ridge-fracture zone-transform intersection by ARCYANA (1975). The second tectonic relationship is required by the abutment of lithosphere that has suffered deformation along the transform fault zone against the spreading ridge axis (Karson and Dewey, 1978; Fig. 6.2.13). The predicted sequence, in the most general terms, consists of the juxtaposition of highly tectonized and faulted rocks, which have passed through the transform domain, with untectonized rocks found at the ridge axis. The contact is an intrusive one, and high temperature post-kinematic metamorphism should affect the deformed rocks of the transform domain close to the contact. The possible intrusive and deformational relationships that can develop as the rocks of the transform domain pass across the end of the basaltic magma chamber of the spreading ridge axis are exceedingly complex, particularly on a small scale, although the general sequence should be that given above. Not only basaltic rocks but also their differentiated and cumulate derivatives may be involved in the complex, overlapping intrusive and deformational events. Karson and Dewey (1978) discuss the process and the resulting relationships in much more detail than is possible here.

Collette (1974) and Turcotte (1974b) have suggested that fracture zones are subject to extension perpendicular to their lengths due to thermal contraction; in addition, Oxburgh and Turcotte (1974) have proposed that suitably oriented cracks, including fracture zones, would be subject to extension because of the change in Earth's radius which would occur if the plate in which they exist moved toward the equator. The amounts and rates of extension produced by either process seem insufficient to provoke significant magmatism or volcanism; no very convincing evidence of



6.2.13 Proposed geologic and geometric development of ridge-ridge transform fault and its fracture zone extensions. (After Karson and Dewey, 1978). **a.** Transform displacement occurs within transform domain (horizontal lines), concentrated for short periods at a principal transform displacement zone (PTDZ). Fracture zone continuations consist of crust with a prior transform history (dashed horizontal lines) against which younger, underformed oceanic crust and mantle are accreted at point M. The contact between these two kinds of crust (X's) is basically intrusive. Numbers 1–8 show position of sections in **b**. **b.** Sketch sections across transform or fracture zone continuation, illustrating geological relationships built up by interaction of horizontal and vertical displacement and magmatism. Half arrows give sense of vertical motion. Positions of sections given on **a**; all views to the right. R. L.—reference level; thick vertical lines—PTDZ; thin vertical lines—limits of transform domain; dashed thin vertical lines—limits of lithosphere with prior transform history; fine stipple—sediments (s); black (p)—lavas; white (SDC) sheeted complex; random dashes (PG)—roof gabbros; oblique lines (MC)—magma chamber; horizontal dashed lines (C)—cumulate gabbros and ultramafics; stipple (UM)—mantle peridotites; vertical wavy lines—peridotite and serpentinite diapirs.

volcanism due to the latter process has been demonstrated.

Pronounced, closed topographic basins at the intersections of spreading ridge axes and transform faults have been ascribed to a magmatic process: viscous head loss (Sleep and Biehler, 1970). These authors have suggested that viscous dissipation due to magma moving upward in a restricted conduit system results in the magma's emplacement at a level below normal isostatic equilibrium. This effect, also proposed to be responsible for the rift along slow-spreading ridge axes, is suggested to be more pronounced at the fracture zone intersections because of the proximity of the older and colder plate edge across the transform-fracture zone.

Volcanism along subaerial transform faults. Where pull-apart basins occur along large strike-slip faults, volcanism may be expected. Examination of such situations—for example, the Salton Sea, the Dead Sea rift and small basins along the North Anatolian fault—reveals that such pull-aparts are not common, that they are usually small relative to the extent of the transform system along which they occur, and that the volume of volcanics directly associated with the pull-apart is also rather minor. It may be that larger volumes of magma occur at depth and that the surface expression is not representative of the amount of magmatism in these pull-aparts; the geothermal activity in the Salton Sea may be evidence of this.

It has been proposed that other volcanic rocks occurring in the general area of some parts of some large continental transform fault systems (for example, the young Arabian basalts, the Hsing-An basalts of China, and some of the young volcanics in the vicinity of the San Andreas Fault) are related to transform-zone tectonics in a less direct way. The evidence that there is a strong connection is less than convincing at the moment; such volcanics are better treated under intraplate volcanism.

6.2.2 Tectonics of non-plate-margin basalt

The relatively small volumes of basaltic igneous material produced in environments other than plate margin environments can be considered from various perspectives. We here make a two-fold distinction between hot-spot basalts and intracontinental rift basalts. No classification of basaltic volcanism is altogether coherent (see, for comparison, section 1.2). Thus there are a number of hot-spot-type volcanic objects on divergent plate boundaries of which the Azores and Iceland are most familiar. Perhaps it is more appropriate to consider the plate boundary as lying on or across these objects than to think of them as representing an unusual kind of plate boundary volcanism.

Hot spot volcanism

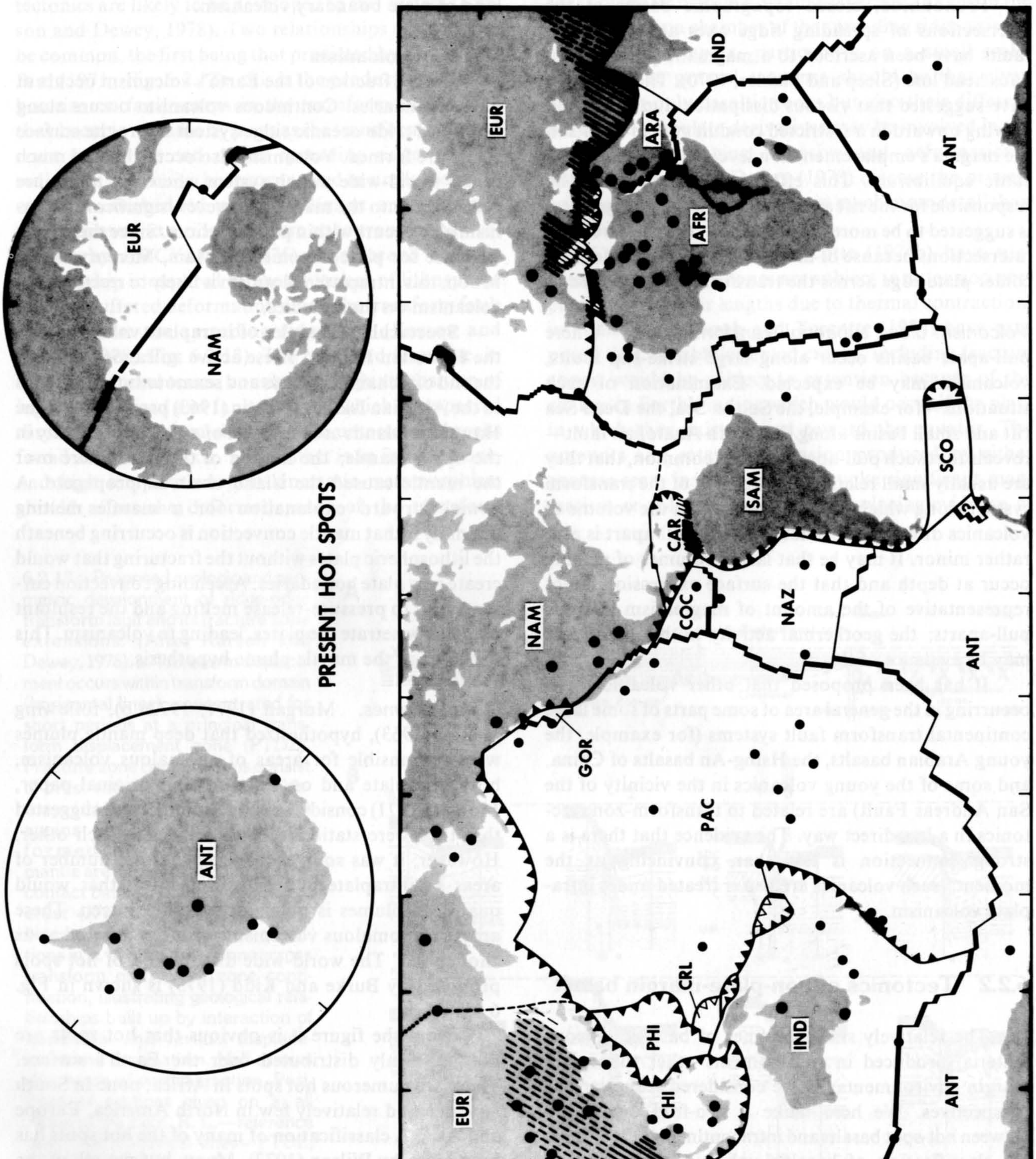
A large fraction of the Earth's volcanism occurs at plate boundaries. Continuous volcanism occurs along the world-wide oceanic ridge system where the surface plates are formed. Volcanism also occurs behind much of the world-wide trench system where the plates are subducted into the mantle. However, significant volcanism also occurs within plate interiors. Since there is no evidence for plate tectonics on Mars, Mercury or the Moon, this intraplate volcanism is likely to resemble the volcanism on those bodies.

Spectacular examples of intraplate volcanism are the Hawaiian Islands. These active volcanoes occur at the end of a chain of islands and seamounts that extends to the Aleutian Islands. Wilson (1963) proposed that the Hawaiian Islands are the result of a melting anomaly in the upper mantle; the motion of the lithosphere over the mantle caused the island chain to propagate. A straightforward explanation for a mantle melting anomaly is that mantle convection is occurring beneath the lithospheric plates without the fracturing that would create new plate boundaries. Ascending convection currents lead to pressure-release melting and the resultant magmas penetrate the plates, leading to volcanism. This is known as the mantle plume hypothesis.

Mantle plumes. Morgan (1971, 1972a,b), following Wilson (1963), hypothesized that deep mantle plumes were responsible for areas of anomalous volcanism, both intraplate and on ridges. In his original paper, Morgan (1971) considered sixteen plumes and suggested that they were stationary with respect to each other. However, it was soon recognized that the number of areas of intraplate and ridge volcanism that would qualify as plumes is much larger than sixteen. These areas of anomalous volcanism are often referred to as "hot spots." The world-wide distribution of hot spots proposed by Burke and Kidd (1975) is shown in Fig. 6.2.14.

From the figure it is obvious that hot spots are not uniformly distributed over the Earth's surface. There are numerous hot spots in Africa, none in South America and relatively few in North America, Europe and Asia. A classification of many of the hot spots has been given by Wilson (1973). Many, but not all, of the hot spots shown in Fig. 6.2.14 are also associated with local crustal upwarping.

One of the assumptions of the deep mantle plume hypothesis is that hot spots move very little with respect



6.2.14 Hot spots (defined as non-plate-margin volcanism) are abundant but not uniformly distributed over the Earth. From Burke and Kidd (1975).

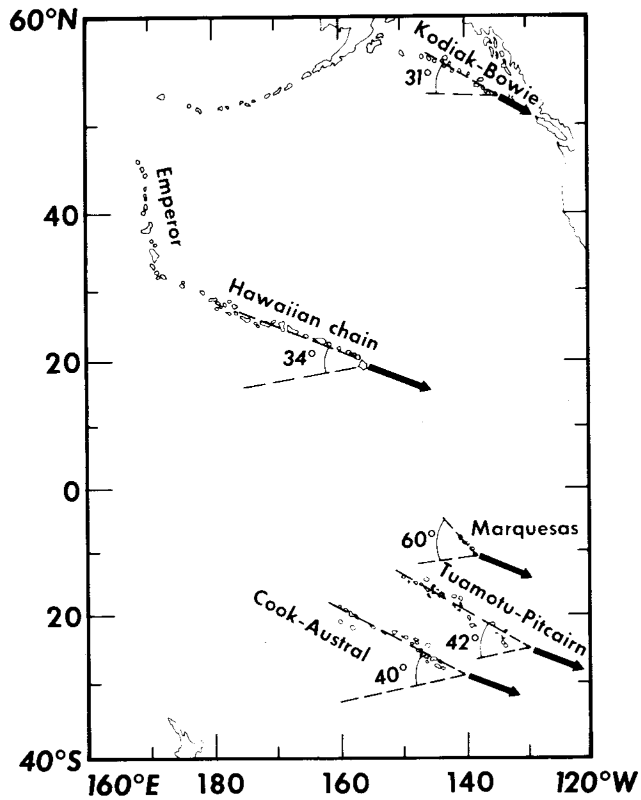
to each other. A number of studies have been carried out to determine whether this is in fact the case. Clague and Jarrard (1973), Molnar and Atwater (1973), Winterer (1973) and Jackson (1976) concluded that the trends of several Pacific island chains can be explained if the plumes migrate with respect to each other at a few millimeters per year. Studies of relative hot spot motions in the Atlantic (Burke *et al.*, 1973; Molnar and Atwater, 1973; Molnar and Francheteau, 1975) also show relative velocities of several centimeters per year. However, since the plate velocities in the Atlantic are also on the order of several centimeters per year, the relative migration of the hot spots is large. In particular, the motion of the Iceland hot spot does not appear to be consistent with the motion of the hot spots in the South Atlantic.

The trends of the Pacific volcanic chains and age progression of volcanism provide the strongest evidence favoring the fixed mantle plume hypothesis. Selected Pacific volcanic chains are illustrated in Fig. 6.2.15. Probably the most comprehensive study of plate motions on a world-wide basis was carried out by Minster *et al.* (1974). These authors inverted the trends of twenty linear volcanic chains and aseismic ridges, utilizing the fixed hot spot hypothesis, and minimized hot spot migration. [Other slowly moving reference frames yield similar results. See, for example, Kaula (1975).] The pole they obtained for the rotation (with respect to a fixed mantle) of the Pacific plate is at 67.3°N and 59.4°W. The angular velocity of the plate is 0.83°/m.y. The predicted directions of propagation of the Pacific Island chains are shown in Fig. 6.2.15. Although the general trends of the volcanic chains are in the predicted directions, significant deviations are seen only if the chains are drawn as lines, rather than as swaths of finite width (Fig. 6.2.16).

In addition to predicting a direction of propagation for volcanic islands, this model also predicts a rate of propagation that may be tested by dating volcanic rocks from islands and seamounts. The data have been summarized by Jackson (1976); ages for the island chains shown in Fig. 6.2.15 are given in Fig. 6.2.17. The age of the volcanic rocks on each island or seamount is plotted against the distance of that island or seamount from the currently active volcanic island at the end of the chain. Also included in Fig. 6.2.17 is the speed of propagation of the chain predicted by the rotation pole and angular velocity given by Minster *et al.* (1974).

The propagation of the Hawaiian-Emperor chain shown in Fig. 6.2.17 is in good agreement with the predicted rate of 8.9 cm/yr. This is even true beyond the change in the trend of the chain about 40 m.y.b.p. Several other chains exhibit similar age progressions, but there are striking exceptions such as the Cook-Austral chain. The mantle plume hypothesis is an attractive means of explaining these Pacific volcanic chains, although without extension and/or modification there are features that remain unexplained, particularly synchronous volcanism along extended lengths of a chain. Most early reports of examples of the latter phenomenon have proven to be erroneous, but some have survived.

The original hypothesis of deep mantle plumes proposed by Morgan (1971) envisioned narrowly confined ($r \simeq 100$ km), cylindrical, thermally convected flows. It was pointed out by Tozer (1973) that the concept of narrowly confined flows is not applicable to high Prandtl-number fluids. Thermal convection at high Prandtl-numbers may result in narrowly confined thermal boundary layers or thermal plumes because of the slow diffusion of heat, but high viscosity leads to a



6.2.15 Some linear volcanic chains of the Pacific with arrows showing directions of propagation about a rotational pole (that of Minster *et al.*, 1974). Dashed lines mark the direction of seafloor spreading that formed the seafloor at the locations of some active hot spot volcanoes and the angles recorded are estimated as separating the trend of the volcanic chains from the local directions of seafloor spreading and fracture zones.

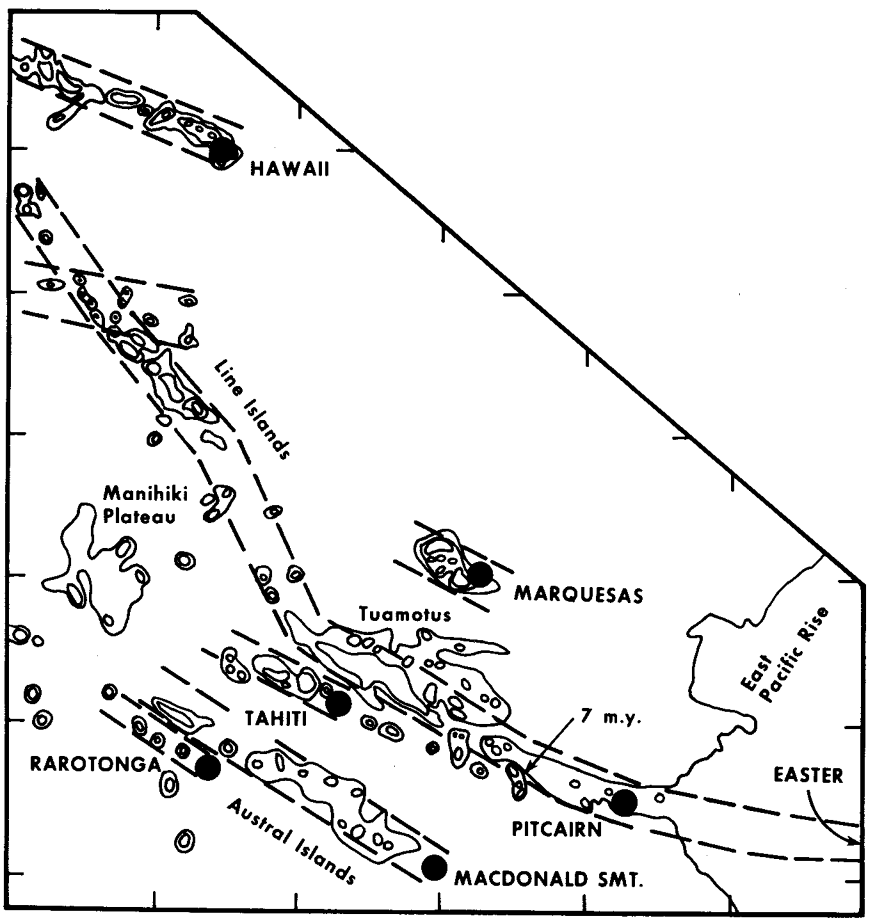
rapid diffusion of vorticity and a diffuse velocity structure.

With an exponentially temperature-dependent viscosity it might be expected that the high temperature region would also be a high velocity region because of the reduction in viscosity. Two-dimensional plumes of this type have been studied by Yuen and Schubert (1977). They find that the flow is primarily within the thermal plume if the temperature difference is sufficiently large. Numerical calculations for the structure of cylindrical mantle plumes have been carried out by Parmentier *et al.* (1975) using an exponentially temperature- and depth-dependent viscosity. These authors found that relatively narrow mantle plumes could be obtained if the convective flow were confined to the upper mantle and if the heating were primarily from below (from the lower mantle). An illustration of this type of shallow mantle plume obtained from the numerical calculations is given in Fig. 6.2.18.

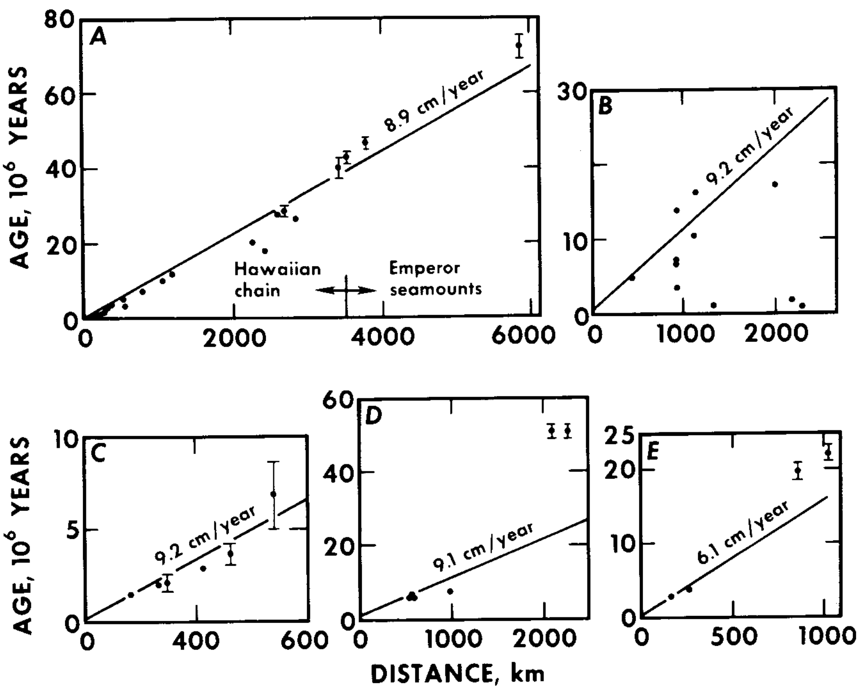
Deffeyes (1972) suggested that deep mantle plumes could be driven by phase changes. A phase change is a local driving force which might drive a local flow; however, it would not induce a narrowly confined, deep mantle plume.

An alternative to thermal plumes is chemical (or compositional) plumes. Anderson (1975) has proposed that surface hot spots result from compositional plumes which rise from the lower mantle. Certainly the diapiric upwelling of lighter material from the lower mantle can result in relatively narrow plumes impinging upon the lithosphere. However, the instability mechanisms that lead to compositional plumes have not been clearly delineated. Also, there is a general depth constraint on all types of mantle convection. If material rises adiabatically from great depths, the degree of partial melting will greatly exceed that which is observed. Anderson's hypothesis implies that the chemical plumes (refractory, radial rods) have been in position since early in Earth's history.

Shaw and Jackson (1973) have suggested that dissipative heating and the consequent partial melting in the asthenosphere produce dense residues that sink in the mantle. The sinking residues create a "gravitational anchor" that fixes the melting anomaly to the mantle.



6.2.16 Another map of Pacific island chains, for comparison with Fig. 6.2.15, illustrating that choice of trends is more realistic when the finite size of hot spots is allowed for, and that two hot spots, at Rarotonga and Macdonald, describe the Cook-Austral track better than one.

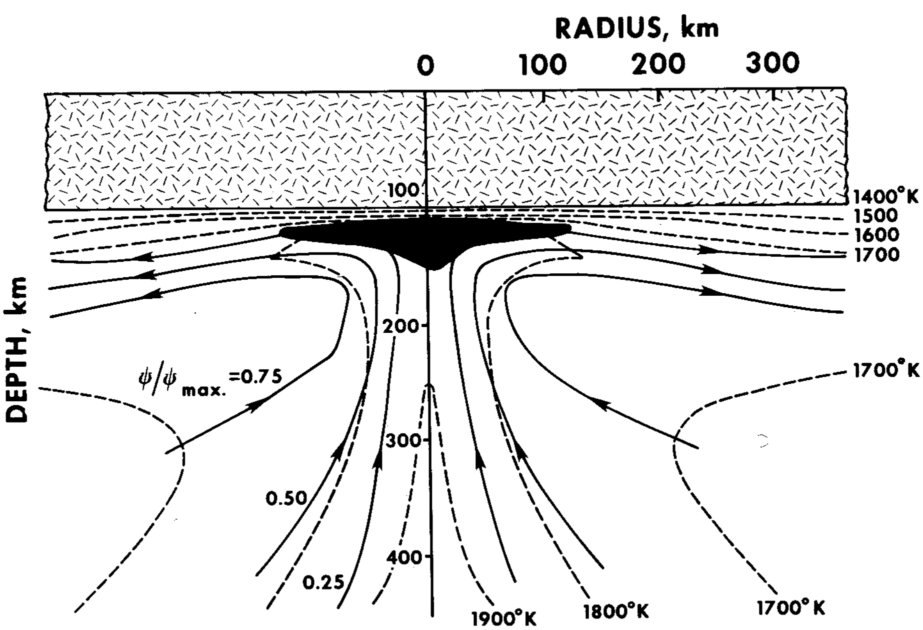


6.2.17 Pacific linear volcanic chains with active volcanoes at one end are shown as aging with distance along the chains. The fit with the rate of propagation for the chains predicted by Minster *et al.* (1974) is good and later work has improved it. The Cook-Austral islands are not readily attributable to a single hot spot.
a. Hawaii-Emperor Chain.
b. Cook-Austral Chain.
c. Marquesas Chain.
d. Pitcairn-Tuamotu Chain.
e. Kodiak-Bowie Chain.

The descending residuum induces further flow and shear heating and melting. This suggestion has yet to be reconciled with the commonly recognized idea that melt and residue are less dense than pristine upper mantle.

The Hawaiian swell is a prominent feature of the topography of the central Pacific. It is a near-circular elevation of the seafloor centered on the Hawaiian Islands. The amplitude of the swell is nearly 1.5 km and its radius is approximately 500 km. There is a long-wavelength positive free-air gravity anomaly of about 15 mgal. The magnitudes of both the topographic swell and of the gravity anomaly are such that they could be associated with a mantle plume beneath Hawaii.

In addition, however, to the many island and seamount chains, some of which we have discussed, there also exists a large number of isolated volcanic islands and seamounts. An example that has been studied in some detail is Bermuda (Pirsson, 1914; Reynolds and



6.2.18 A lithospheric plate 100 km thick (dashed ornament) is underlain by a rising shallow mantle plume. Continuous lines are flow lines (normalized values) and dashed lines are isotherms. Black is zone of partial melting (Parmentier *et al.*, 1975).

Aumento, 1974). A suite of younger volcanics has an Early Oligocene age of about 38 m.y. and an older group an age of 60–110 m.y.; the seafloor beneath Bermuda has an age of about 110 m.y. Bermuda is not part of a volcanic chain and is difficult to explain by the mantle plume hypothesis. It is interesting to note that, like Hawaii, Bermuda is associated with a lithospheric swell. The amplitude of the swell is smaller (approximately 750 m) than the Hawaiian, but the radius is similar. Since Bermuda is moving with respect to the mantle, a mantle plume which had produced volcanism at Bermuda 38 m.y. ago would not be centered at Bermuda today. One preferred interpretation is that, if Bermuda's elevation is due to a mantle inhomogeneity, this inhomogeneity is not that which was responsible for the 38-m.y.-old volcanism.

Important sources of evidence for the existence of deep or shallow mantle plumes are seismic studies. Seismic studies can be used to search for velocity anomalies and/or regions of seismic attenuation in the mantle. An anomalous zone of seismic velocities and attenuation has been reported in the vicinity of the core-mantle boundary beneath Hawaii by Kanasewich *et al.* (1972, 1973) and by Kanasewich and Gutowski (1975). A number of deep mantle inhomogeneities, including one beneath Hawaii and several others not associated with surface hot spots, was reported by Julian and Sengupta (1973). However, the interpretation of these data has been questioned by Wright (1975) and Green (1975). These authors suggest that the observed anomalies are due to upper mantle inhomogeneities beneath the receiving seismic arrays in North America. At present, the seismic evidence for deep mantle plumes beneath hot spots is ambiguous at best. Further verification of the velocity anomaly beneath Hawaii is required. Deep anomalies beneath other hot spots must be sought and some are being found: for example, those in Yellowstone (Hadley *et al.*, 1976) and at Trinidad (Okal and Anderson, 1975).

Ellsworth and Koyanagi (1977) have reported on travel delays associated with teleseismic arrivals at 21 stations on Kilauea Volcano in Hawaii. They found no velocity anomalies at depths greater than 30 km. If any type of mantle convection were occurring beneath surface hot spots, a zone of anomalous mantle would be expected, such as that which occurs beneath ocean ridges and beneath back-arc spreading centers. The failure to find anomalous mantle beneath intraplate hot spots such as Hawaii suggests that these hot spots are not associated with special mantle features.

Lithospheric fractures. Although some fraction of intraplate volcanism may be associated with convective processes, it is difficult to explain all intraplate volcanism in this way. It is recognized that some intraplate volcanism is associated with extensional tectonics; examples are the East African Rift, the Rhine Graben, and the Basin and Range Province in the western United States.

An alternative hypothesis for intraplate volcanism is that magmas from the asthenosphere penetrate the lithosphere where tensional fractures occur. Betz and Hess (1942) proposed that the Hawaiian Islands are the result of a propagating fracture. Jackson and Wright (1970) and Green (1971) discussed the petrology of the Hawaiian Islands in terms of magmas rising through a cracked lithosphere. Turcotte and Oxburgh (1973, 1976) discussed the geometry of lithospheric fractures and the sources of the required stresses.

In order for tensional failures of the lithosphere to occur, elastic stresses must build up over long periods of time. The presence of an outer swell due to the bending of the lithosphere under the load of the Emperor seamounts (Watts and Cochran, 1974) indicates that elastic stresses are not relaxed for at least 40 m.y. The failure of the lithosphere under extension is not likely to resemble the tensional failure of a rock in the laboratory. Turcotte and Oxburgh (1973) have suggested that the lithosphere will fail plastically according to the failure criteria given by Bijlaard (1935). A thin, malleable plate under tension will fail plastically at an angle

$$\theta = \tfrac{1}{2}\left(\cos^{-1}\tfrac{1}{3}\right) = 35°16'$$

where θ is measured from a line perpendicular to the direction of the tensional stress. However, other workers have thought that it is inappropriate to apply membrane theory to terrestrial stresses.

The trend of the Hawaiian chain lies at an angle of 34° with respect to the direction of seafloor spreading defined locally by the magnetic anomalies and fracture zones (Fig. 6.2.15). If this angle is associated with the plastic failure angle θ, the conclusion is that the tensional stress causing the plastic failure of the Pacific plate in the vicinity of the Hawaiian Islands is nearly perpendicular to the direction of seafloor spreading. Stresses with this orientation that are large enough to cause failure of the lithosphere could result from thermal contraction of the oceanic plate (Turcotte, 1974a).

The rate of propagation of the tensional fractures is not predicted directly by the hypothesis. If thermal stresses cause the tensional failures, then it might be expected that the failure would propagate in the direction of seafloor spreading at about the seafloor spreading velocity. Taking the seafloor spreading velocity to be 6 cm/yr, the propagation velocity would be $6/\cos 35° = 7.3$ cm/yr.

Seismic studies are also relevant to the propagating fracture hypothesis. Focal mechanism studies of major earthquakes can give information on the state of stress which led to the seismic rupture. On June 27, 1962, a magnitude 6.1 earthquake occurred near the Kaoiki fault system on the island of Hawaii. This earthquake has been studied by Koyanagi *et al.* (1966). They have concluded that the earthquake was due to either right-lateral strike-slip movement on a fault trending N25° E or left-lateral strike-slip movement on a fault trending N65° W. The latter direction follows the trend of the island chain, the former direction is roughly parallel to the Kaoiki fault trace. If this earthquake were due to a simple shear failure of the brittle lithosphere under tension, the direction of the tension would be N15° W. This is within two degrees of the direction of tension (N13° W) given by the plastic failure hypothesis.

On April 26, 1973, a magnitude 6.2 subcrustal earthquake occurred beneath the northeast coast of the island of Hawaii. First motion studies reported by Koyanagi *et al.* (1976) suggest that the earthquake was due either to right-lateral strike-slip movement on a fault trending N30° E or to left-lateral strike-slip movement on a fault trending N70° W. If this earthquake were due to a simple shear failure of the brittle lithosphere under tension, the direction of the tension would be N20° W. Despite being at different depths and different locations, the focal mechanisms and the direction of the inferred tensional stress for the two earthquakes are in remarkably good agreement. They are also in good agreement with the predictions of the plastic failure hypothesis.

An essential part of the lithospheric fracture hypothesis is that the fracture allows an essentially artesian flow of magma from the asthenosphere to the surface (see section 6.6). Normal lithosphere behaves as an impermeable cap rock, but under tensional failure the hydrostatic head drives the magma to the surface. The physical processes by which magmas are transported through the lithosphere are poorly understood. However, Weertman (1971) has argued that magma in the asthenosphere can nucleate a crack in the lithosphere, provided that the lithosphere is in tension. The magma then follows the propagating fracture to the surface. Anderson and Grew (1977) have suggested that magma migrates through a lithospheric plate in tension due to "stress corrosion." Solomon and Sleep (1974) have attributed seamount chains to the propagation of the plate relative to the intraplate stress field maintained at the boundaries of the plate. Fujita and Sleep (1978) have extended the hypothesis to small seamounts associated with ridge-transform intersections. At present, these hypotheses are difficult to test because the intraplate stress is poorly determined. They also suffer from the inability to account for all, or even many, hot spots.

Africa. As shown in Fig. 6.2.19, Africa is the site of extensive intraplate volcanism. Burke and Wilson (1972) argued that Africa is stationary with respect to the present mantle convection pattern; therefore, mantle plumes could produce fixed volcanic centers on the continent. However, the world-wide studies of Minster *et al.* (1974) concluded that Africa is moving towards the northeast at about 2 cm/yr. This is within the limits of error predicted by the Burke and Wilson hypothesis.

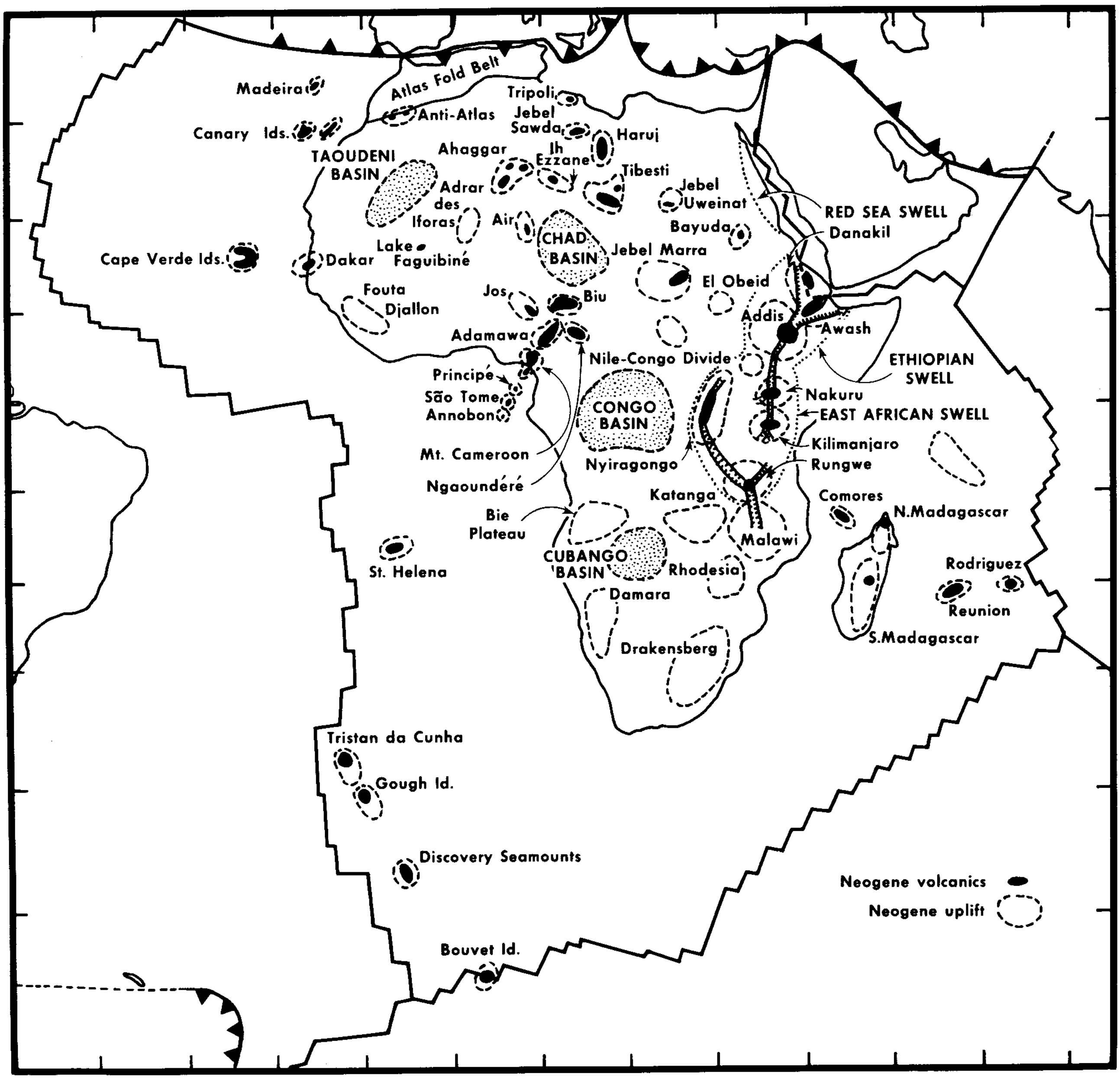
The intraplate volcanism of Africa is associated with several different tectonic settings as shown in Fig. 6.2.19. Much of the volcanism is related to the East African Rift system. Burke and Wilson (1976) have associated the East African Rift with seven hot spots. These hot spots are defined by crustal doming, which is an important aspect of the rift system. Oxburgh and Turcotte (1974) have suggested that the East African Rift is formed by tensional failure of the African plate due to membrane stresses, but supporting evidence is not persuasive. They have argued that the doming and volcanism are secondary features resulting from the tensional failure.

Studies of the focal mechanisms of earthquakes in southern Africa have been carried out by Maasha and Molnar (1972) and by Scholz *et al.* (1976). These authors conclude that an extensional failure of the lithosphere is occurring in southern Africa and that the East African Rift system is propagating southward along a preexisting zone of weakness.

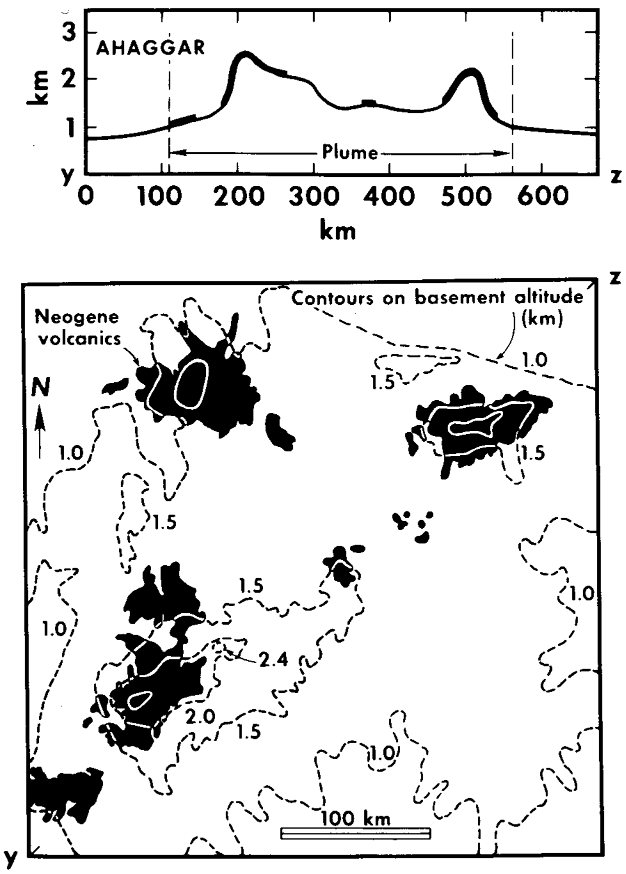
Other areas of extensive recent volcanism in Africa are the Tibesti area in northeast Chad and in Ahaggar to the west (Fig. 6.2.20). In addition to the volcanism, these areas are domed. North of Tibesti, the Haruj volcanics are associated with apparent doming of the basement. Adjacent to the continental margin in the Gulf of Guinea is the Cameroon Line. This is a nearly linear series of active and recently active volcanic centers, which passes from oceanic onto continental crust.

Farther to the north on the continental margin of Africa there is the recent volcanism of the Canary Islands. This is a group of volcanic islands similar to the Cape Verde Islands farther to the south. The association of the Canary Islands with a propagating fracture has been discussed by Anguita and Hernan (1975), but the evidence for this association is not persuasive. Such volcanic centers, which lie near the continental margins of the South Atlantic, may be associated with large stresses caused by changes in crustal thickness (Bott and Dean, 1972); however, there is not a great concentration of volcanic centers in this tectonic position, particularly on a global scale.

Western United States. The western United States is an area of extensive, young volcanism. The distribution of Quaternary (< 1.5 m.y.b.p.) and Pliocene (1.5–7 m.y.b.p.) volcanics is given in Fig. 6.2.21. Since the San Andreas fault is recognized as a major plate boundary between the Pacific and North American plates, the volcanism of this area may be classified as plate margin volcanism. However, it extends over a width of 1500 km. The spreading direction as defined by the Blanco Fracture Zone is not parallel with the trend of the San Andreas fault. This difference has led Atwater (1970) to conclude that active subduction is occurring along the Oregon-Washington coast. This subduction explains the line of active andesitic volcanoes extending from Mt. Baker in northern Washington to Mt. Lassen in northern California, although the lack of seismic activity associated with the hypothesized subduction zone must be noted.



6.2.19 Hot spots on the African plate are associated with Neogene uplifts and with the Neogene and active East African rift system. Note the numerous uplifts without volcanism and the intracontinental basins lying between the uplifts.



6.2.20 Map and cross section (after Black and Girod, 1970) illustrating that uplift is the dominant property of the Ahaggar hot spot and that Neogene volcanics, mainly basalt flows, cap the most elevated areas.

The distribution of seismicity in the western United States is shown in Fig. 6.2.22. The main area of active volcanism, seismicity and tectonics is bounded on the east by the Rio Grande Graben, and on the north by the Snake River Plain, although there is considerable seismicity in the northern Rockies. The Basin and Range Province is an extensive zone of horst and graben tectonics, high heat flow, hot springs and recent volcanism. Focal mechanism studies (Smith and Sbar, 1974) indicate the entire area is in tension with the directions indicated by the arrows in Fig. 6.2.22.

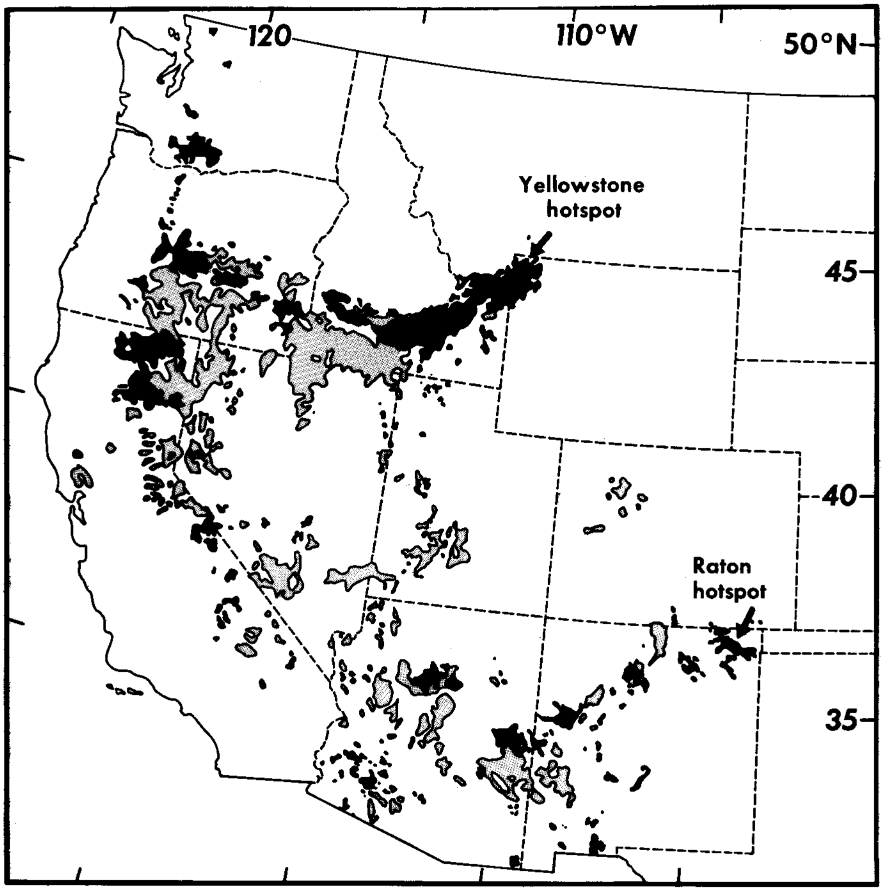
Atwater (1970) and Christiansen and Lipman (1972) have attributed the tensional tectonics and the volcanism of the western United States to forces exerted at the San Andreas plate boundary, with tensional failures of the lithosphere permitting the flow of magma to the surface. Scholz *et al.* (1971) have related the volcanism and tectonics of the western United States to behind-arc spreading.

Morgan (1971) has associated one of his original 16 plumes with the extensive recent volcanics of the Snake River Plain. If such a plume or hot spot is postulated in the vicinity of Yellowstone caldera at the eastern end of the plain, the trend of the eastern section of the plain is in good agreement with the direction of propagation predicted by Minster *et al.* (1974). A large number of volcanic rocks in the Snake River Plain have been dated by Armstrong *et al.* (1975). The ages of these (and other) rocks as a function of the distance from the Yellowstone caldera are given in Fig. 6.2.23. Also included is the rate of propagation of the volcanics predicted by Minster *et al.* (1974) on the basis of the fixed hot spot hypothesis. Although the age of the oldest volcanics tends to increase with distance from Yellowstone, the agreement with the predicted rate is rather poor. It is also evident from Fig. 6.2.23 (as well as Fig. 6.2.21) that there are recent volcanics along much of the length of the Snake River Plain. It is difficult to associate extensive basaltic volcanism over a distance of 500 km directly with a hot spot beneath Yellowstone. Smith *et al.* (1974, 1977) have suggested that the Snake River Plain volcanics have followed a preexisting zone of weakness and that a wedge-shaped fracture is propagating eastward in response to intraplate deformation.

Suppe *et al.* (1975) have attributed much of the volcanism of the western United States to two hot spot tracks. In addition to the Yellowstone hot spot, they have proposed that a hot spot is located near Raton, New Mexico. This is shown in Fig. 6.2.21. They would extend this hot spot track to the White Mountains of eastern Arizona and possibly as far as the Gulf of California. These authors have conceded, however, that these hot spots cannot explain all the intraplate volcanism of the western United States. Burke and Wilson (1976) have located four hot spots in the western United States (see Fig. 6.2.14). Even this number of hot spots cannot explain all the volcanics of the area.

It was shown by Gruntfest (1963) that the heat produced by viscous dissipation in a fluid with a strongly temperature-dependent viscosity could lead to an instability. Anderson and Perkins (1974) have suggested that this type of thermal runaway could be responsible for magma generation in the asthenosphere and the type of volcanism observed in the western United States.

Eurasia. Much of the active volcanism over this huge continental area can be related to extensional features developed in association with the Alpine-Himalayan collision (see section 6.2.1; Molnar and Tapponnier, 1975). The identification of volcanic areas related to mantle heterogeneity in a continental area where much



6.2.21 Quaternary (filled) and Pliocene (open) volcanic areas in the western United States. The Yellowstone and Raton hot spots of Suppe *et al.*, 1975, are located by arrows.

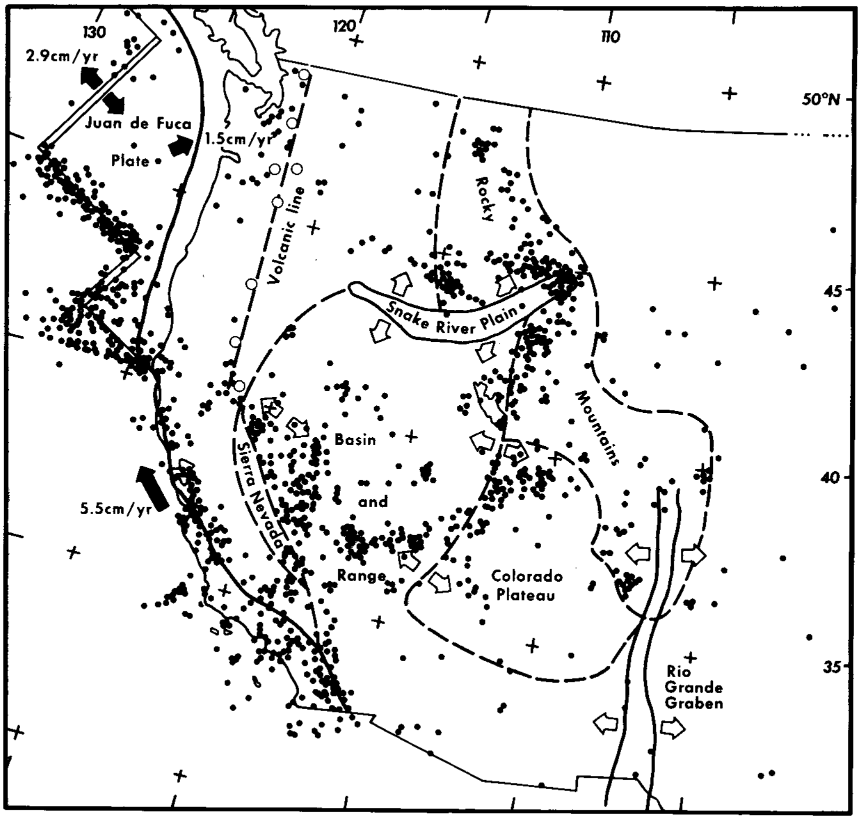
of the volcanism is associated with continental collision remains a major problem. Distinction between the two kinds of basaltic activity may be extremely difficult; it may prove necessary to resort to geochemical indices.

Volcanism in continental rifts

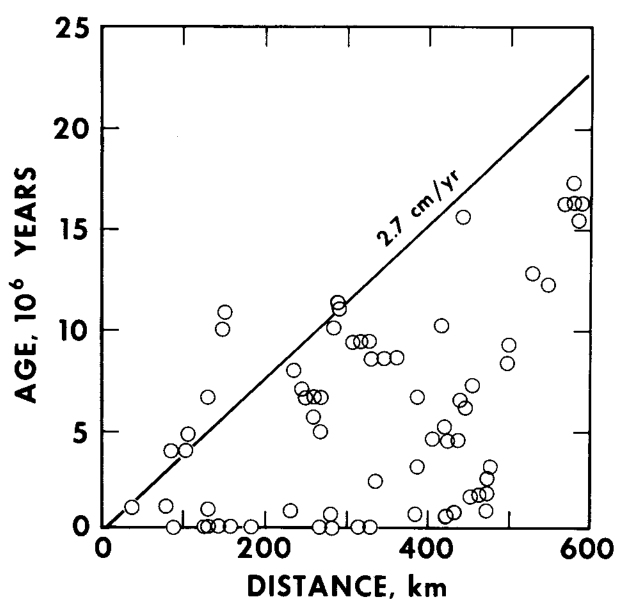
The word "rift" is used in geology in a variety of ways, most of which have nothing to do with major structures. Gregory (1921) was the first to use and later to popularize the word rift for a large-scale structure when he described the East African Rift. He defined a rift valley as a long depression between parallel normal faults. Here a modified version of Gregory's definition is used and **rifts** are defined as "elongate depressions overlying places where the entire thickness of the lithosphere has ruptured in extension." This omits reference to parallel faults because many familiar rifts (for example, the Connecticut Rift) have a major fault on only one side. In some rifts it is hard to be sure whether a boundary is a fault or a steep monoclinal flexure. By referring to "the entire thickness of the lithosphere," emphasis is given to rifts as large-scale structures and small depressions are excluded. The reference to extension is to distinguish rifts from the other major fractures, such as transforms and thrusts at convergent plate boundaries, that penetrate the whole lithosphere.

Rifts are the most common major lithospheric fractures. Since rupture of the lithosphere is involved in rift formation, they are most readily developed where the lithosphere is thinnest—that is, in the oceans. This section is solely concerned with continental rifts, which are now—and probably always have been—much less abundant than oceanic rifts (see section 6.2.1).

Two observations have been gleaned from continental rift studies made over the last ten years: first, intracontinental rifts are very numerous; second, rifts have developed within continents in a range of structural environments. These are all environments with dominantly extensional tectonics. In this section, active and ancient rift occurrences associated with a number of different extensional plate-tectonic regimes are described. It is necessary for us to define the Wilson cycle, which is the concept that permits application of plate tectonic principles to the record of historical geology within the continents.



6.2.22 Earthquakes of the western United States (shown as filled circles) superimposed on a map showing geological provinces. Open circles are the Cascade province volcanoes. Solid arrows are directions of motion across the Juan de Fuca ridge and along the San Andreas fault. Open arrows are directions of motion inferred from seismic focal mechanisms. (Smith and Sbar, 1974).



6.2.23 Ages of volcanic rocks on the Snake River Plain as a function of distance from the Yellowstone Caldera (Armstrong *et al.*, 1975). The solid line is a rate of propagation for a Yellowstone hot spot as predicted by Minster *et al.*, (1974). One simple interpretation of this figure is that volcanism started as the hot spot passed a locality, and continued in relation to subsequent rifting.

Rifts in the Wilson cycle. Our discussion of basaltic volcanism in rifts comes twelve years after Tuzo Wilson first drew attention to the best way in which to apply the tenets of plate tectonics to studies of continental geology. Wilson (1968a) argued, before the American Philosophical Society, that plate tectonics showed that oceans are opening in some places on the surface of the Earth and closing in others. He suggested that the history of the Earth's surface, therefore, can be considered as a record of the opening and closing of oceans. Since ocean floor disappears by subduction or by erosion during or after obduction, the record of these cycles has to be sought within the continents. Looking for records of these cycles is the most powerful way of interpreting continental geology. Wilson analyzed the cycles in terms of oceanic evolution from youth (continental rupture) to old age (continental collision). Dewey and Burke (1974) later suggested that complex interwoven cycles of ocean opening and closing of the kind recorded within the continents be called Wilson cycles.

Rifts at continental breakup. Rifts occur at all stages of the Wilson cycle because extensional tectonics can develop locally at all stages; however, those associated with continental rupture are both particularly well-developed and particularly accessible to study. The East

African Rift system is the best known of continental rift systems. Progress in its understanding has accelerated with appreciation that its present activity dates only from the beginning of the Neogene, when the African plate appears to have come to rest with respect to the underlying mantle convection pattern (Burke and Wilson, 1972). Distinction of the current episode of rifting from a very similar episode in Africa 100–200 m.y. ago that was associated with the break-up of Gondwana has proven to be important. Rift studies that define the timing of events accurately have become of particular importance in recent years, and seem likely to remain important in the future. It will only be possible to establish how rifting relates to other tectonic events if the relative timing of the phenomena is accurately known.

Two other rift properties that have been firmly established in the East African Rift system are the following: rift faults commonly follow and reactivate old structures (McConnell, 1974); rift igneous rocks are almost wholly mantle-derived. Crustal material seems to be involved only to a very small extent in igneous rocks of the East African and nearly all other rifts. Detailed studies of igneous rocks associated with the East African Rift system have emphasized the widespread occurrence of alkaline rocks, most of which appear to have last equilibrated with the mantle at depths of 60–100 km, and many of which appear to have suffered complex subsequent histories. Comparable alkaline rocks occur in many rift systems, but so do tholeiitic rocks, suggesting equilibration at much higher mantle levels.

Teleseismic studies have been among the most fruitful of geophysical investigations in the East African Rift system. These studies have shown the peculiar nature of the mantle below the rift system as well as the extensional character of most rift earthquakes. Although local seismic networks have yielded important results, the very limited number of modern seismic stations on the African continent has proven to be something of a barrier to research. Gravity studies are well advanced in East Africa and have proven most susceptible to analysis where information from them has been coupled with other data. Seismic refraction studies, for example, have indicated the presence of high velocity, dike-like objects in some rifts, which has helped control gravity interpretations.

If the East African system continues to propagate and eventually becomes a system extending across the African continent, then a new ocean may develop and the African plate may split in two. If this happens, many presently active rifts will be left in one or another of the two new continents. Such rifts will have failed to develop into oceans. Failed rifts of this kind, stretching into continents at a variety of angles from Atlantic-type ocean margins, are very numerous and are rapidly becoming the best-known of all fossil rifts (Burke, 1976a). Our knowledge of these environments is expanding because the sediment fill of the failed rifts is oil-bearing in some places. Because petroleum exploration, seismic reflection studies, and wells can yield accurate information on how rifts subside with time over intervals of tens of millions of years, characterization of the thermal behavior of individual rifts or parts of rifts may prove possible. The widespread occurrence of failed rifts, trending at various angles to Atlantic continental margins, may be responsible for the difficulty of precisely locating the boundary between continent and ocean in many areas.

Although the general behavior of failed rifts at Atlantic margins seems to follow a regular pattern, there are exceptions. For example, the South Atlantic opened about 120 m.y. ago, and many failed rifts associated with this event have passed through normal sequences of events and became filled with marine sediments in the Aptian. However, the Rio Salado and Colorado rifts (Urrien and Zambrano, 1973) continued to accumulate nonmarine sediments until much later (about 80 m.y. ago), and the Benue Trough in Nigeria behaved like a small ocean that appears to have opened and shut over the interval between 125 and 80 m.y. ago (Burke and Dewey, 1974).

Convergent boundary rifts. Rift developments at Andean continental margins and in island arcs are most prominent at places where the lithosphere is thinnest, that is, along the line of the volcanoes themselves (see section 6.2.2). These rifts, which are known in such active Andean arcs as New Zealand and Sumatra-Java, achieve their greatest tectonic significance only when the volcanic arc splits along them to form a marginal basin as happened, for example, in the Japan Sea at the beginning of the Miocene (Sillitoe, 1977). Rifts of Basin and Range type appear to be associated with imperfect transform motion along continental boundaries, but few fossil examples have been recognized as yet.

Collisional rifts. Perhaps the greatest variety of rift development is associated with continental collisions. Failed-rift systems formed at rupture become reactivated as rifts striking into the collisional mountain belts, forming the population of objects that Shatskiy (1947) called **aulacogens**. Also, a whole new set of rifts, of which the Lake Baikal and Rhine rifts seem the best active examples, are set up as a result of intracontinental strains established during the collision. These objects have been called **impactogens**, and they can be readily distinguished from aulacogens where their geological history is well-established.

In addition to these two major classes of collision-related rifts, final closing of an ocean is accompanied by very complex rift production (Tapponnier, 1977). This can be seen very well in the Mediterranean, where formation of the rifts of Corsica and Sardinia 20 m.y. ago accompanied rotation of these islands away from France (Dewey *et al.*, 1973), and where formation of the Aegean rifts within the last 20 m.y. has been related to the westward motion of Turkey with respect to the Black Sea and the Mediterranean (Dewey and Şengör, 1979). Rift development during the closing of the Mediterranean has been considered by Tapponnier (1977), who concluded that all of the rifts represent secondary extensional phenomena. If the closure of the Mediterranean continues, rifts of the Aegean and Corsican-Sardinian types are likely to be compressed out of all recognition. Such structures are probably preserved only as obscure areas within mountain belts that show signs of late-stage extension before final compression. For this reason, rifts of this kind are of less general interest in continental geological history than the four major rift classes (continental rupture rifts, failed rifts at Atlantic margins, aulacogens, and impactogens). A generalization of this type always has exceptions and the Devonian rift basins of Norway (Horn, Solund and Hornelen) form just such an exception (Steel, 1976). They appear to have formed within the Caledonides at continental collision and closely resemble the Thakkola rift of the high Himalaya today.

(a) *Aulacogens.* Although the aulacogen concept is over thirty years old, widespread use of the term among scientists in the United States dates only from attempts to interpret the aulacogen concept in plate tectonic terms (Burke, 1977). Shatskiy (1947) used the word only for rifts striking into fold belts, and it seems unwise to use it, as has sometimes happened, for all ancient rifts that have failed to develop into oceans. Shatskiy also emphasized that the depositional and structural history of the aulacogen begins at the same time as the depositional history of the fold belt into which it leads, thus eliminating the structures now being called impactogens. An unfortunate fact in the history of the use of the word "aulacogen" is that one of Shatskiy's type examples, the Dnieper-Donets structure, is a failed rift that strikes not into a fold belt but into the North Caspian depression. The North Caspian depression is an area which is probably underlain by ocean floor formed in the Devonian that has been covered with a great thickness (~14 km) of sediments but has, so far, escaped both subduction and obduction (Kidd, pers. comm., 1975).

Although aulacogen histories can be very diverse, it is often possible to make a two-fold division: an earlier continental rupture and ocean opening phase with development of typical rift features, including alkaline igneous rocks, basal coarse clastics and higher evaporites; and a distinct, later post-collisional phase with development of coarse clastic sediments derived from the collisional mountain belt, and of strike-slip faulting. The outstanding structural features of many aulacogens include the enormous thicknesses of sediments and volcanics that they contain. Shatskiy (1947) pointed out that the upper member of an aulacogen sediment pair commonly extends over an area wider than that of the initial rift. This seems a particular case of the more general observation that numerous major intracontinental basins, such as the Chad, Paris and Michigan basins, overlie older rifts (Burke, 1976b).

(b) *Impactogens.* In map view an impactogen resembles an aulacogen because it also is a rift structure striking into a fold belt. The distinction between the two is made initially by ascertaining whether rift history goes back as far as the opening of the ocean whose closure is represented in the fold belt. If it does not, and if rifting dates only from the time of ocean closure in the fold belt, then the rift is an impactogen. For example, both the Rhine graben and the Polish trough strike from northern Europe into the Alpine fold belt. The start of the geological history of the former is contemporaneous with the mid-Eocene Meso-Alpine collisional event (Şengör *et al.*, 1978), while that of the latter started in the Triassic. Thus, the Rhine graben is an impactogen while the Polish trough is an aulacogen.

Although the Rhine graben reaches to the edge of the Alpine System in the Jura, impactogenal rifts associated with the Himalayan collision (the Baikal and Shansi grabens) develop far from the main collision zone. Molnar and Tapponnier (1975) have suggested that these rifts formed as consequences of the eastward escape of China from crushing between the vise-like jaws of India and Asia. The Baikal and Shansi rifts are linked to the main collisional mountains in Tibet and the Himalaya through a system of strike-slip faults and folds. Timing of the Himalayan collision is not yet as well-established as that of the main Alpine collision, so it is hard to know how closely the start of faulting in Shansi and Baikal coincides with the collision marked by the Indus Suture.

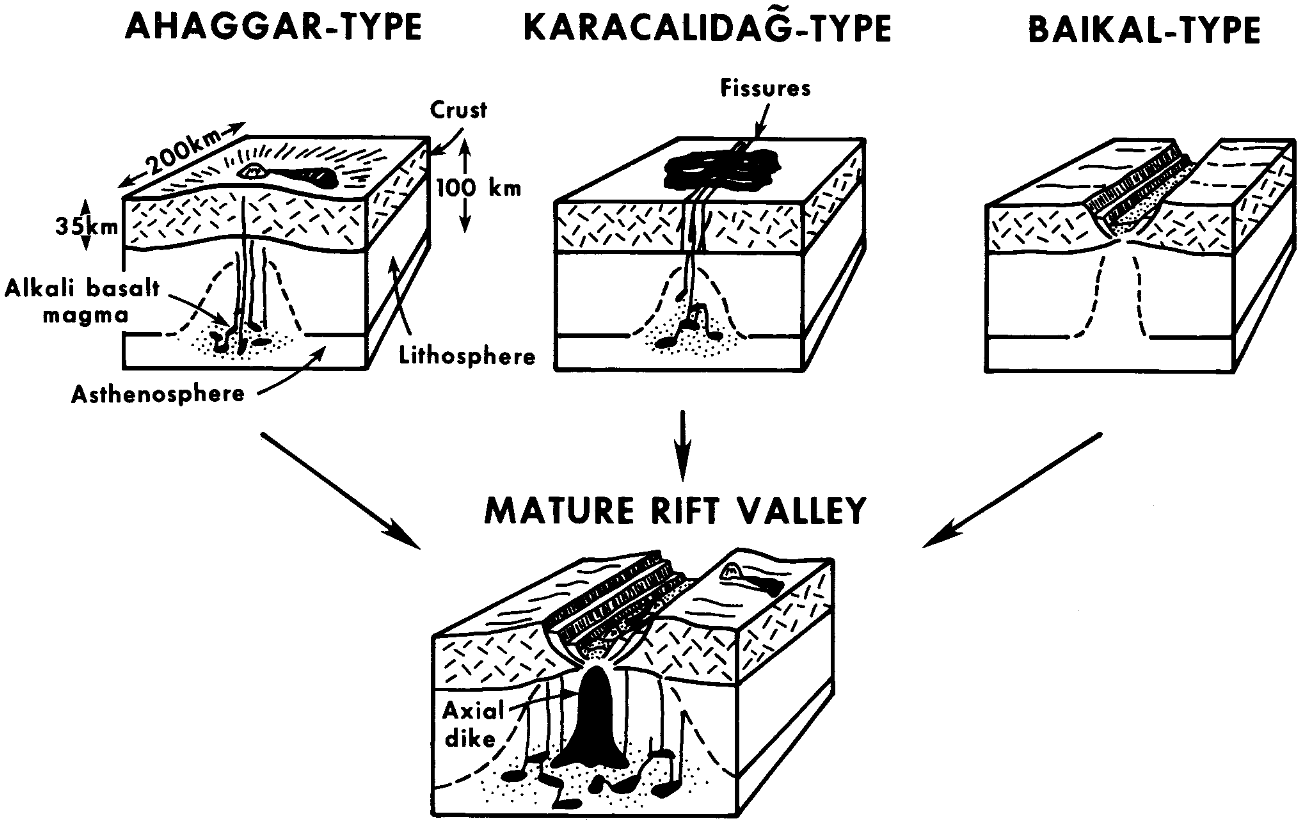
Impactogens seem particularly well-developed in association with the Hercynian fold belt of Europe. The Hercynian ocean appears to have opened in two stages: one in medial Devonian times along a rift extending from western Britain to the Urals, and the other in Early Carboniferous times along a rift with a very different strike in Iberia. The timing of the Devonian event indicates that the rift may have been formed as an impactogen related to collision in the Caledonides. The second rift could have been an Acadian impactogen (Burke and Sawkins, 1977). Although evidence of the formation of the Hercynian ocean is very incomplete, evidence of its closure at the beginning of the Permian is much stronger. Rifts developed over much of northwestern Europe at the beginning of the Permian (the Oslo graben being the best known such rift). These rifts are strong candidates for interpretation as impactogens. The rifts of northwest Europe established in this Permian event have been repeatedly reactivated in Triassic, Jurassic, Cretaceous and Cenozoic times (Whiteman *et al.*, 1975). This is most obvious in the North Sea where geological development of the huge oil province has been dominated by rift reactivation (Woodland, 1975). A possible explanation of renewed subsidence in a long inactive rift, so typical of the North Sea, is that a thermal pulse in the mantle induces transition from a basaltic to an eclogitic mineralogy in a sub-rift axial dike system.

Relative timing of rifting and volcanism. Basaltic volcanism is a feature common to most rifts. In continental rifts, with which we are concerned here, these basalts are mainly alkaline. The relative timing of basaltic volcanism with respect to the associated rifting shows considerable variation. Currently, there are two main hypotheses to explain the origin of basaltic magmas beneath rifts. One favors the preliminary cracking of the lithosphere due to differential stresses resulting from two-dimensional plate evolution e.g., membrane stresses: Turcotte and Oxburgh, 1973; and Turcotte, 1974a; stresses due to the collision of continents: Molnar and Tapponnier, 1975; and Şengör, 1976; stresses resulting from the propagation of existing accreting plate margins into continents: McKenzie and Weiss, 1975 . This cracking upsets the temperature/pressure balance of the underlying mantle, resulting in its partial melting. The partially melted area experiences a volume increase that may induce a post-rifting uplift of the overlying lithosphere, such as appears to have happened in the Upper Rhine Graben (Şengör *et al.*, 1978). The other hypothesis suggests that a complicated convection pattern in the mantle is responsible for doming and cracking the lithosphere, thereby giving rise to extensional fractures and eventually to rifts (Burke and Whiteman, 1973; Burke and Dewey, 1973b). In the former view, the expected sequence of events is rifting-doming-volcanism, whereas in the latter it is doming-volcanism-rifting (Fig. 6.2.24). It is our contention that both hypotheses are compatible with the available data, but are applicable to fundamentally different tectonic environments. The first hypothesis is applicable to regions where rifting is the result of horizontal movements of plates and their interaction, and the mantle plays a passive role. The second hypothesis seems to be valid where (small-scale?) convection in the mantle (McKenzie and Weiss, 1975) directly affects the overlying lithosphere and induces rifting as a combined result of primary vertical tectonics (uplift) and possibly later horizontal motion (initial spreading). It appears that this latter mode of rifting occurs mainly on plates that are fixed with respect to the underlying mantle, as Africa has been suggested to have been since 25 m.y. ago (Burke and Wilson, 1972). At present, the former mode of rifting is by far the more widespread of the two.

Passive mantle hypothesis. The horizontal movement of plates and their interaction give rise to differential stresses within the lithosphere that may result in rifting. Turcotte and Oxburgh (1973) and Turcotte (1974a) argued that membrane stresses can be generated within a plate moving longitudinally on the surface of the Earth, due to the ellipticity of the Earth, and this may result in rifting. Burke and Dewey (1974), however, indicated that the elastic stresses thus generated cannot be stored for long periods of time and therefore may not be as important in generating rifts as Turcotte and Oxburgh (1973) suggested. Moreover, the supporting evidence that Oxburgh and Turcotte (1974) used to apply this idea to the East African rifts consisted of incorrect K/Ar data and misinterpretation of paleomagnetic results.

Molnar and Tapponnier (1975) and Şengör (1976) pointed out that continental collision, especially of irregular margins, also generates extension within continents, resulting in rifting more or less perpendicular to the direction of convergence. If collision-induced strike-slip faulting also causes rift formation, as is the case for the Baikal rift (Sherman, 1978), such rifts are not likely to have any systematic orientation with respect to the convergence direction. In the foreland of the Alps in Europe, for example, rifting events are well-correlated with the Mesoalpine collision (Trümpy, 1973; see Şengör, 1976, Table 1). Here rifting appears to have predated major basaltic volcanism. In the Upper Rhine Graben, rifting started during medial Eocene with some minor volcanism of 42 m.y. age along the master faults (Şengör *et al.*, 1978). Major basaltic volcanic activity began at the northern end of the graben about 25 m.y. ago (Lotze, 1974). This volcanic center, the Vogelsberg, appears to have continued its activity until very recently. In the south, well within the major graben trough, the Kaiserstuhl volcano began its activity about 18 m.y. ago; this, however, unlike the Vogelsberg, was a short-lived event. In the Lower Rhine Graben there are few volcanics, but the maars of Eifel appear to be related to Plio-Pleistocene northwest-trending faults (Greiner and Illies, 1977). In Bohemia, both along the Thuringian disturbance and in the graben between the Bohemian and Thuringian blocks (*Schollen*), volcanism began during the Oligocene and lasted through the Pliocene; rifting along these lines had begun in the earliest Tertiary and accelerated during late Eocene (Lotze, 1974). It appears that in Europe it is this kind of inter-Scholle activity that caused basaltic volcanism which is largely of olivine-nepheline-bearing alkaline type.

Western Turkey and the Basin and Range Province of the western United States are areas of extensive rift development, characterized by numerous subparallel grabens in "rift clusters." Both areas are related to large, imperfect strike-slip plate boundary zones (western Turkey: Dewey and Şengör, 1979; western U.S.: Atwater, 1970). In western Turkey, rifting began during the late Miocene and is currently active. Most rifts here



6.2.24 Sketches illustrating how a mature rift valley may evolve from three different origins. In the Ahaggar-type, the mantle is active and uplift and volcanism precede rifting. In the Baikal-type, the mantle is passive and rifting is not preceded by doming or volcanism. The Karacalidağ-type illustrates a complexity: horizontal extension forms fissures through which magma wells up prior to major down-faulting and rift formation. The sequence of events documented in the geological record may make a Karacalidağ-type rift look very like an Ahaggar-type rift, although the mode of origin more closely resembles that of a Baikal-type rift. However rifts originate, their structures develop in such a way that all mature rift valleys look rather similar as illustrated in the bottom figure. Because of this similarity, working out rift origins depends on a good record of geological history.

are devoid of volcanics with the exception of the Gediz and Simav grabens (Dewey and Şengör, 1979). On the northern shoulder of the Gediz graben, an area 50 km long and 20 km wide is covered with Pliocene to recent alkaline basalts that are nepheline-, leucite- and hornblende-bearing (i.e., kulaite: Washington, 1894; Erinc, 1970). These basalts emanated from fissures that can be shown to be controlled by faults related to the Gediz graben. Zeschke (1954) documented a similar history for the Simav basalts.

In the Basin and Range Province major rifting related to the present regime began about 18 m.y. ago (Noble, 1972) and was closely followed by basaltic/rhyolitic bimodal volcanism (McKee *et al.*, 1970). The geographic extent and the possible close temporal association with a strike-slip regime (Hamilton, 1970) of the western Siberian rift system of Triassic age (Logatchev, 1977) suggest that it might be a fossil analog of the Basin- and Range-type rift regimes, where generally basaltic volcanism postdates rifting.

Active mantle hypothesis. Rifts active on the African Plate can be interpreted as products of relatively simple mantle-lithosphere interaction. The control on rifting appears to be the lack of relative motion between the African Plate and underlying convective circulation over the last 25 m.y. (Burke and Wilson, 1972; Thiessen *et al.*, 1979). The sub-African Plate mantle-lithosphere interaction is manifested in several phenomena: the distinctive basin and swell structure; the occurrence of intraplate volcanism (largely on swell crests); the development of rifts and the evolution (in the Red Sea and in the Gulf of Aden) of rifts into oceans (see Burke, 1977, for a review).

Throughout southern Africa, swell crests carry no volcanoes and rifts are mainly not on swells but are reactivated old structures. This indicates that a sequential development from volcanoes on swells to rifts such as Burke and Whiteman (1973) distinguished is not universally recognizable. Further evidence that rift development is complex can be seen in the active rifts of East Africa. Volcanism there is very unevenly distributed along the length of the rifts. In some areas (for example, near Addis Ababa) volcanism is dominantly tholeiitic, whereas in other areas it is dominantly alkaline (Baker *et al.*, 1972). The implication of this observation is that the petrology of the igneous rocks is a very poor indicator of rift style compared with more direct structural/stratigraphic features such as topography, faulting and sediment fill. The general absence of signs of interaction between magma and continental crust in the rift igneous rocks is perhaps the most significant feature and is to be expected in a regime where axial dikes and extension dominate.

Older episodes of rifting induced by similar mantle interactions are generally hard to identify, but the lack of motion between Africa and the spin axis during the break-up of Pangaea has been taken as evidence that African rifting during that time was induced by this interaction (Burke and Dewey, 1973b). The Pangaean-rupturing rifts on either side of the Atlantic, which formed just before that ocean opened, provide further examples of the diversity of styles of volcanism. Some Pangaean rifts appear to be without volcanic material; others are associated with extensive pre-rift igneous activity; others show igneous activity only when the rifts are well-developed. As is evident in East Africa at present, compositional diversity is the rule. In some areas carbonatites and alkaline syenites abound (e.g., Los, Bagbe, Songo), whereas in others tholeiitic basalts are the only igneous rocks. It seems that terrestrial rifts may develop in a variety of ways, and only in exceptional circumstances can the record within individual rifts be deciphered in sufficient detail to permit clear understanding of all the processes that have operated in their development.

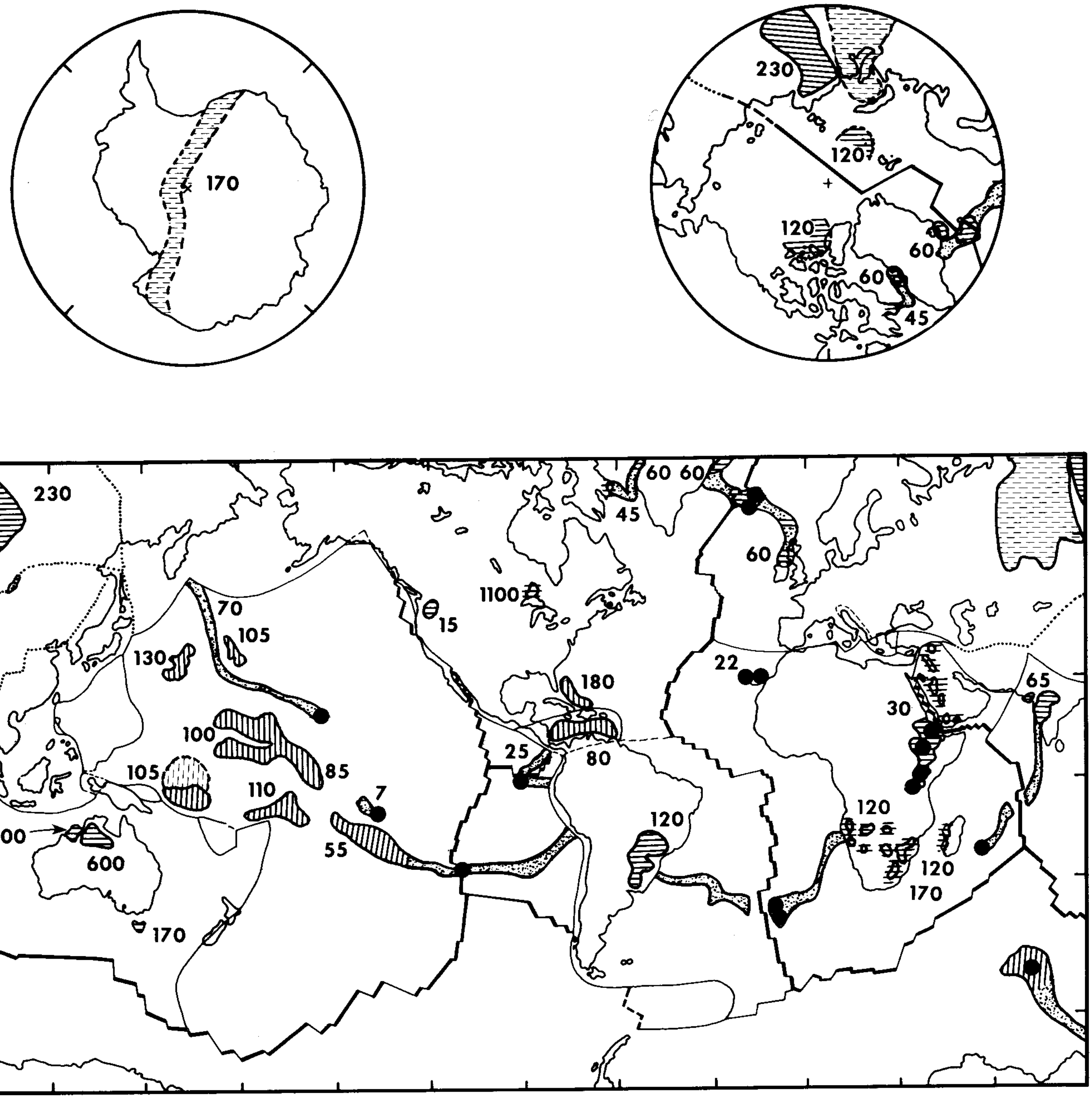
Flood basalts. Young volcanism not related to plate boundaries is widespread (Burke and Kidd, 1975, 1980); although in some ways there is great variety in its expression, there is sufficient similarity between the occurrences of non-plate-margin volcanism that it seems appropriate to treat them in a single category. Many areas of active intraplate volcanism are well-exposed in Africa and have been well-studied; the association of volcanics with structurally and topographically high areas is well-established (Burke and Whiteman, 1973; Thiessen *et al.*, 1979). The volcanics of Dakar, which are located near sea level, actually are not exceptions to this observation since subsurface data (Spengler and Delteil, 1966) reveal an underlying youthful structural elevation. There is a complete spectrum of expression of intraplate volcanic areas in Africa, from uplifts without any volcanism on them (e.g., Fouta Djallon) through minor volcanism (e.g., Aır) of alkaline type to more abundant alkaline volcanism (e.g., Tibesti), to areas of voluminous tholeiitic flood basalts together with comparatively minor alkaline volcanics (e.g., Ethiopia). In all of these cases there is a structural uplift associated with the volcanic area; the size of this uplift is somewhat variable although most tend to be 100–200 km across (Burke and Whiteman, 1973). There is perhaps a tendency for the areas with larger volumes of volcanics to have larger diameter uplifts, although it is possible to resolve subsidiary uplifts within the larger ones (e.g., Ahaggar; Fig. 6.2.20). Thus it is not wholly clear if the larger structures are merely groups of smaller ones, or whether the smaller diameter uplifts in them are secondary to the larger (but generally lower amplitude) uplifts.

This spectrum of intraplate volcanism can be seen, although less well-displayed, on other continents. More significantly, it is also developed in the oceans and the alkaline and tholeiitic volcanics in oceanic hot spots are essentially indistinguishable from those in continental hot spots. Once again the African plate provides many of the best examples. Oceanic intraplate volcanic areas also show a variation in volume from regions like Hawaii, with abundant tholeiite and minor alkaline volcanics, to relatively small edifices like Ascension, that are mainly alkaline volcanics. Very small volume intraplate volcanic areas, like some of those on the African continent, are harder to detect in the oceans, and uplifts without accompanying volcanism are even more difficult to detect without making direct attempts to find them (Menard, 1973).

Because alkaline magmas are erupted in places in present island arcs and collisional orogens, it is not yet possible to detect "intraplate type" or "hot spot type" activity within such zones of convergence even though it may well occur there. However, it is possible to detect such volcanism along divergent plate boundaries because of the alkaline character of some of the magmas, excess volumes of magma, and the associated structural uplift. The occurrence of hot-spot-type volcanism at discrete and long-lived sites along divergent plate boundaries is one of the most significant properties of their distribution and helps considerably to winnow the many hypotheses put forward for their origin. In particular, since the lithosphere is thin to virtually nonexistent at spreading ridge crests and is being created progressively away from them, it is unlikely that the crack propagation hypotheses of Turcotte and Oxburgh (1973) and Oxburgh and Turcotte (1974) and the dense anchor-asperity hypothesis of Shaw and Jackson (1973) are correct. An "active mantle" hypothesis for hot spot origin, such as that proposed by Wilson (1963), is more satisfactory.

A similar spectrum of size and volume of magmas can be seen in those hot spots sited on divergent plate boundaries. The central and north Atlantic contain the best examples. These range from the anomalously-elevated area on the ridge at 45° N where there seems to be relatively little excess volcanism, through the Azores with mostly alkaline volcanism constructing small islands above sea level, to the extreme of Iceland with its voluminous excess tholeiitic volcanism and, in relative terms, minor alkaline magmatism. Anomalous (with respect to spreading ridge basalt) geochemical signatures (see section 7.4.2), particularly of $^{87}Sr/^{86}Sr$ values (White *et al.*, 1976), correlate exactly with discrete anomalously elevated ridge crest areas (in Iceland, 45° N, Azores, and near 34° N at the Colorado Seamount). The distinctive geochemistry of the "excess" magmas argues strongly for a separate (deeper) source for them (Schilling, 1973) compared with normal spreading ridge basalts. It is because the latter are produced at most places in fairly constant amounts through a wide range of spreading rates that the "excess" character of the hot spot magmatism along ridge axes can easily be detected.

The connection between flood basalts and the largest hot spots is well shown by Iceland (Burke and Kidd, 1980). The island itself consists of flood basalts. The hot spot tracks that lead away from Iceland go to large areas of flood basalts in East Greenland and the northern British Isles (Fig. 6.2.25) that were erupted during the initial rifting that led to successful seafloor spreading in the northern Atlantic beginning about 60 m.y. ago. The relics of associated central alkaline volcanoes of that age are well-known from Scotland. A younger but similar situation is seen in the Afar (Fig. 6.2.25) with the Ethiopian Traps, which erupted before and up to the opening of the Red Sea and the Gulf of Aden. If the Pacific plate were not moving so rapidly with respect to the source of the Hawaiian magmas, the accumulation of igneous material at present rates of production would clearly rival Iceland. Although the Hawaiian-Emperor hot-spot trace (Fig. 6.2.25) does not lead back to a site of rifting and a large flood basalt pile, it is appropriately grouped with other flood basalt-producing objects. Several other large hot spots have tracks leading back to rifting sites and to accumulations of flood basalts (Fig. 6.2.25). In particular, Tristan/Gough leads to the Kaokoveld and Paraná basalts, erupted just before and up to the opening of the South Atlantic 120 m.y. ago, and Reunion leads back to the Deccan Traps erupted about 65 m.y. ago just before rifting in the Gulf of Khambhat and removal of the Seychelles from India (McKenzie and Sclater, 1971). The Galapagos hot spot, although its tracks do not lead back to preserved flood



6.2.25 Flood basalts on Earth. There is a gradation between hot spot volcanism (only the larger, active hot spots are shown as black circles), which may cover only a very small area, and very large areas of flood basalt and sill intrusion. Flood basalts on continents (horizontal lines, numbers indicating age of extrusion in m. y.) are linked to hot spots which are either adjacent to them, as in the case of the Ethiopian traps, or along ocean floor tracks (stippled). Extensive basaltic areas, especially sill complexes (vertical lines), are common in the Pacific Ocean, but are also known in the Atlantic and Indian Oceans. None of these is older than 180 m.y., because older oceans have disappeared. Flood basalts are the most extensive of non-plate-margin igneous phenomena on Earth. (Dashed horizontal lines—possible extension of flood basalts mostly or entirely in subsurface; dashed vertical lines—possible extension of submarine sill complex.) From Burke and Kidd (1980).

basalts, is sited on a spreading ridge that started about 25 m.y. ago by rifting across older oceanic crust generated at the East Pacific Rise (Hey, 1977). It may not be a coincidence that the Galapagos hot spot, as shown by its area and the volume of magmatic products in its tracks, is a relatively large one, like others that have been listed as associated with flood basalt production and initial ocean opening. Ocean opening in the Labrador Sea-Baffin Bay was accompanied by flood basalt volcanism about 60 m.y. ago which now is seen on Disko Island and Cape Dyer of Baffin Island. The hot spot that produced this volcanism then produced the shallow ocean floor of Davis Strait as a track, but this hot spot is now obviously extinct. Perhaps it is also not coincidence that spreading ceased in the Labrador Sea-Baffin Bay at the same time or shortly after the hot spot died. This case illustrates the point emphasized by Vogt (1972), that the volumes of magma generated in hot spot sites vary with time, although we do not think there is strong evidence for synchroneity in this variation among many hot spots. Thus, the amount of magma produced by the Hawaiian hot spot, judged by the volume of the track, was very small at the time of and shortly after the bend formed in the Hawaiian-Emperor chain. It may have been no more volumetrically impressive at that time than one of the smaller present ocean island hot spots, such as St. Helena. The points we wish to make here are that voluminous flood basalt-producing hot spots are the same kind of object as the volumetrically smaller ones, and that one can change into the other and back again with time. They may also die out, in terms of their volcanic expression, temporarily or permanently. They vary a great deal in the length of time during which large quantities of tholeiitic magma have been produced, from the short burst of the Columbia River basalts (Baksi and Watkins, 1973) and Deccan Traps (Wellman and McElhinny, 1970) to the more extended histories of Iceland or Hawaii.

Large flood tholeiite events in the oceans seem, in some instances, to produce huge sill complexes, probably because of the limited abilities of even mafic lava to travel underwater before chilling. Such sill complexes (Fig. 6.2.25) underlie large portions of the Caribbean (Burke *et al.*, 1978) and the submarine plateaus of the western Pacific (Winterer, 1976). The Mid-Pacific Mountains, one of the latter, may perhaps be traced to the hot spots of either Easter Island and/or Pitcairn Island through the Tuamotu-Line Islands track. Sill complexes are, of course, also important components of many continental flood basalt events, particularly the early Jurassic Karoo and Ferrar dolerites of South Africa and Antarctica, and the Mid-Cretaceous Isachsen diabase of the Canadian Arctic (Fig. 6.2.25).

It is of note that all major flood basalt events, where at least some of the extensive lavas and/or sills are still preserved, were connected with extensive rifting and, in most cases, with successful opening of an ocean. Two large and well-known ones not associated with successful ocean opening are the Columbia River basalts and the Siberian Traps.

6.2.3 Tectonics of active volcanism elsewhere in the solar system

Tectonics of active volcanism on Io

Elsewhere in the solar system, active volcanoes have been recognized only on Io, the innermost of the Galilean satellites of Jupiter (Morabito *et al.*, 1979). The density of Io is such that these volcanoes could be erupting basalt as well as the more volatile sodium and sulfur-rich products that have been recorded.

The occurrence of what appear to be calderas on at least one volcano on Io indicates the existence, at least locally, of a rigid lithosphere, since calderas on Earth are formed by fault collapse. The surface of Io appears to be highly mobile although no coherent large-scale plate structure has emerged. In this respect the tectonic style of Io may be similar to a style that existed on other objects of planetary dimensions early in the history of the solar system and that probably existed only during the period of the high impact flux. The reason for the intense tectonic activity on Io appears to be that jovian tides generate very large amounts of energy within Io which is dissipated as heat (Peale *et al.*, 1979). Tidally induced tectonics do not appear to be significant elsewhere in the solar system although tidal influences triggering volcanic eruptions have been recognized on Earth (e.g., Mauk and Johnston, 1973).

An apparent peculiarity of the tectonics of Io is that, although the tidally induced volcanic activity seems likely to have been in progress for a large proportion of the history of the satellite, today's activity still involves large amounts of volatile material. There does not appear to have been on Io, as on the Earth and the Moon, irreversible fractionation of low density material to make outer parts like the lunar highlands and the terrestrial continents, oceans and atmosphere.