

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

These possible ocean floor fragments in greenstone belts with associated mafic volcanics and volcanoclastics closely resemble those dismembered ophiolites interpreted as being of marginal basin origin in Phanerozoic mountain belts (e.g., Bird *et al.*, 1971; Dalziel *et al.*, 1974). Burke *et al.* (1976) suggested that it is very likely that the contents of marginal basins compose a large

proportion of Archean greenstone belts, and that there are two reasons why any Archean oceanic crust that may be identified will be more likely to represent marginal basin rather than main ocean crust. First, the Archean ocean with many scattered microcontinents and arcs appears likely to have had many more areas in which young ocean floor was close to arc and microcontinental sediment sources than in later times. Second, marginal basin ocean floor has a greater chance of being young (and hence thin and hot) near to a subduction-resistant arc or remnant arc (microcontinent) than does main ocean floor, and therefore has a greater chance of being obducted and preserved.

Summary of Archean tectonics

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

6.5 TECTONICS OF BASALTIC VOLCANISM ON OTHER TERRESTRIAL PLANETS

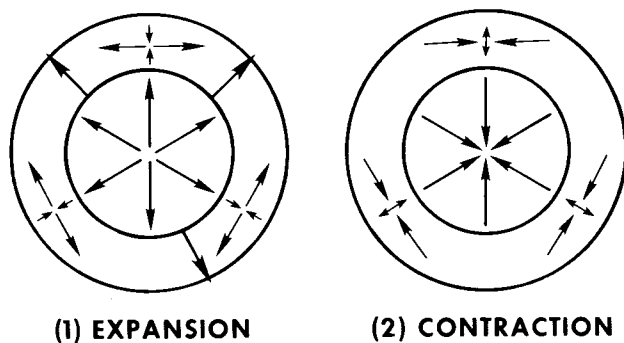
Mars, Mercury and the Earth's moon share the property of having at present a stable global lithosphere. This is in contrast to Io, where lithospheric mobility is so great that all lithosphere is probably

recycled frequently (on a time scale perhaps less than 10^7 years), and to the less extreme, plate-structured Earth. On Earth, oceanic lithosphere is fully recycled on a time scale of about 10^8 years, but most material that accreted

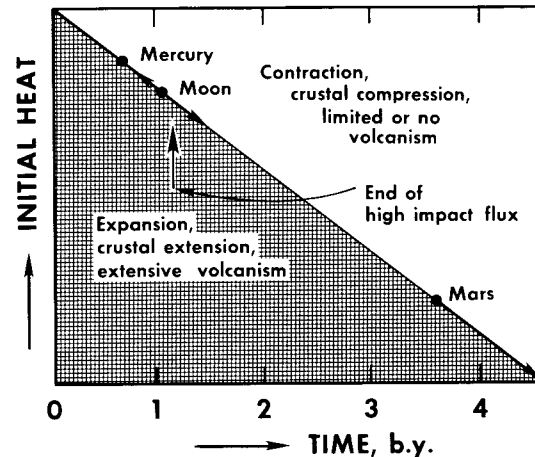
to continents remains, forming part of the record of Earth history. Mars, the Moon and Mercury show no sign of comparable lithospheric mobility after the end of the period of heavy bombardment because, at least on part of their surfaces, they preserve a record of the great impacts. Other, presumably younger, surfaces show no morphologic evidence of lithospheric recycling. Data from Venus are sparse, but the evidence points towards a lack of present-day plate tectonics.

Solomon (1978) has reviewed the tectonic style of planets with immobile lithospheres ("one-plate planets") and has suggested that, in general, after they become one-plate planets, they may record on their surfaces an early stage of expansion and a later stage of contraction (Fig. 6.5.1). In the former stage extensional features, especially normal faults, and igneous activity are general, while in the latter stage compressional structures and an absence of igneous activity (eruption being inhibited by compressive stresses) are characteristic. The Moon and Mercury have passed through the first stage and are now in the second, while Mars was in the earlier stage until relatively recently, and it is not clear whether it has yet entered the second (Fig. 6.5.2).

Volume change alone results in an isotropic distribution of stress in the lithosphere and should produce randomly distributed tectonic features (folds, faults, fracture sets). Superimposed on this secular and spherically symmetric stress state are regional stress systems associated with thermal inhomogeneities in the mantles of these one-plate planets. Closely related are stresses due to thermally induced variations in lithospheric thickness. The mantles of these planets generate basaltic



6.5.1 Sketch illustrating the behavior of a planet with a continuous immobile lithosphere. An earlier warming phase is associated with expansion, surface igneous activity and normal faulting of the lithosphere (annular region not shown to scale). Later cooling is associated with shrinking and absence of igneous extrusion. If the lithosphere fractures in this phase it does so in compression. Based on Solomon (1978).



6.5.2 One-plate planets go through a secular evolution from interior expansion, crustal extension, and widespread volcanism to interior contraction, crustal compression, and cessation of volcanism. The time of transition from expansion to contraction is a function of the magnitude of initial extensive heating, shown here with an arbitrary scale. The transition, shown by the diagonal line, marks the approximate peak of planetary volume. The peak volumes (with uncertainties) are shown for Mercury, Moon, and Mars. Note that only Mars went through the transition after the end of the high impact flux. Modified after Solomon (1978).

magmas at various stages of their history. When and where these magmas reach the surface are functions of the lithospheric stress history, specifically of the environments for extensional tectonics, which in turn are dependent on the superposition of regional and global stress systems. Additionally, basaltic magmas reach the surface through fractures created by large-scale impact events. However, a favorable stress history is required to keep these fractures open between the time of impact and the time of magma generation.

In addition to mantle thermal inhomogeneities, other phenomena concentrate tectonic stresses. These include lithospheric stresses due to loading by volcanic material, polar wander of the lithosphere, tidal despinning and, for the Moon, orbital recession. The relative importance of these various mechanisms varies among the planets and also during the evolution of an individual planet.

6.5.1 Tectonics of basaltic volcanism on the Moon

Introduction

On the Moon, tectonics (large-scale lithosphere evolution) has been overwhelmingly dominated by

impact-related phenomena. Almost the entire evolution of the lunar lithosphere took place during the high impact flux prior to 3.9 b.y. ago (for reviews, see Taylor, 1975; Guest and Greeley, 1977). Subsequent events involved only the flooding of the maria 3.9 to 3.65 b.y. ago (Lucchitta and Watkins, 1978) and later, small-volume eruptions lasting until about 3.0 b.y. ago (Waserburg *et al.*, 1977).

The contrast between the Moon and Earth is almost complete. Tectonics on Earth is readily related to the long-term thermal evolution of the interior of the planet (see Chapter 9), and active tectonics persists to this day. The role of impact in terrestrial tectonics is generally regarded as minimal (but see, for contrasting views, Seyfert and Murtagh, 1977; and Weiblen and Schulz, 1977). On the Moon, impact dominance is so great that unrelated tectonism cannot be distinguished (Muehlberger, 1977). What we have learned about lunar thermal behavior and its variation with time has come from indirect inferences from the secular variation in structural response to both impact and subsequent volcanic emplacement (Solomon and Head, 1980a). In particular, the gravity ("load") tectonics associated with basalt infilling of the large impact basins is decipherable and places constraints on the global stress system associated with lunar expansion and contraction. However, there is no unequivocal direct evidence for global stresses large enough to have caused widespread lithospheric failure comparable to the global distributions of scarps on Mercury (see below).

Lunar highlands

The lunar highlands are inferred, from petrological and geochemical evidence, to represent material formed by global melting early in lunar history (see section 1.2.10). The highlands have been so severely modified by impact that no structural or tectonic evidence that could help to show how this process operated has yet been distinguished. The general elevation of the lunar highlands above the level of the basins (Kaula *et al.*, 1973) is evidence of crustal buoyancy, but the obvious analogy between buoyant highlands above lunar basins and buoyant terrestrial continents above ocean basins cannot be pursued too far. Highlands once covered much, if not all, of the lunar surface, whereas it is unlikely that continents on Earth ever were more extensive than they are now. Although ocean basins are also younger than the neighboring continents, the processes by which ocean basins form are quite unlike those that created the younger mare basins within the lunar highlands.

Early Earth-based observations (Fielder, 1961, 1965; Baldwin, 1963) identified an extensive, supposedly

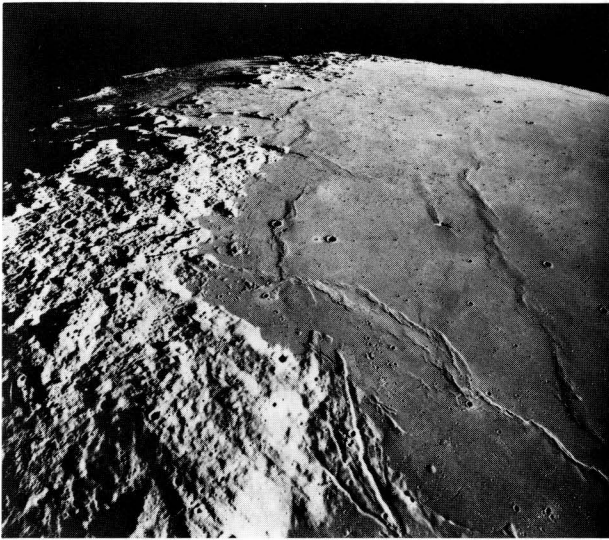
systematic, global distribution of lineaments that has been collectively referred to as the lunar grid. Similar lineament systems have been recognized by some on Mercury and Mars and in the past have been suggested for the Earth as well. The lack of distinction, or ignorance, of the relative ages of development of included features (such as scarps, ridges, crater walls) makes it difficult to interpret or assess the tectonic importance of these supposed systems. The inevitable inclusion in these systems of a large proportion of lineaments generated in various ways by impacts leads us to consider the tectonic significance of such "grids" as minimal.

No direct correlation can be seen between ancient lunar structural features and highland volcanism. Highland volcanism on the nearside shows evidence for structural control related to the Imbrium Basin ring system (Head and McCord, 1978; Hawke *et al.*, 1979) or adjacent mare infilling (Head and Wilson, 1979). Structural and volcanic features become increasingly rare toward the central farside (Scott *et al.*, 1977), with mare deposits concentrated in topographically low areas (Schultz, 1976; Stuart-Alexander, 1978).

Whitford-Stark (1974) and Bryan *et al.* (1975) drew attention to the pronounced fractures and evidence for volcanic processes that have modified large lunar impact craters. Many such craters are located within the lunar highlands close to the mare-highlands boundary, although a significant number also occur within highland regions far from such boundaries (Schultz, 1976). Floor-fractured craters may mark the locations of igneous intrusions that have uplifted and fractured the crater floor. Not all mare-highland boundaries show examples of fractured crater floors, but the mechanism for their formation is interpreted to be the intrusion of magmas within the fractured crustal material close to major impact basins. A similar mechanism has also been proposed by Head and Wilson (1979) to explain the occurrence of dark-halo craters close to basin margins.

Mare basins

The mare basins are large, basalt-flooded, impact-generated structures up to 2000 km across that have modified the lunar highlands. The distribution of basins across the lunar surface is not uniform because they are the products of only a few tens of random impacts. Consequently, the concentration of basins on the lunar nearside is not attributable to any large-scale structural heterogeneity within the Moon. Linear rilles and ridges (see Chapter 5; Fig. 6.5.3) are structures associated with mare basins that have been considered tectonic in origin, but "most rilles and ridges can be ascribed to subsidence of the mare basin region in response to the



6.5.3 A view of the western edge of Mare Serenitatis showing both linear rilles (lower center) and mare ridges (center and right). Note that linear rilles are flooded by later mare basalts. Apollo 17 metric photo 0953.

load produced by the basalt fill" (Solomon and Head, 1980a), and they are thus the result of gravity tectonics rather than internally generated tectonic phenomena.

The role of the high-density basalt fill (mascons) in the circular maria as a means of loading the lunar lithosphere has long been considered important (e.g., Baldwin, 1968; Phillips *et al.*, 1972) and structural features within maria (Fig. 6.5.3) often have been attributed to basin loading and subsidence (e.g., Baldwin, 1968; Muehlberger, 1974; Runcorn, 1974; Scott *et al.*, 1978; Maxwell *et al.*, 1975; Maxwell, 1978).

Solomon and Head (1979) made an ingenious tectonic inference from the distribution and age of these features. They pointed out that both linear rilles and ridges were formed before 3.6 b.y. ago (Lucchitta and Watkins, 1978), while only ridges continued to form after that time, persisting until at least 3.0 b.y. ago. Such a chronology was interpreted to be due to a global stress system that was superimposed on local mare basin stresses. This global stress field was extensional prior to 3.6. by ago and compressional after that time, inhibiting the formation of the linear rilles during the final 600 m.y. of lunar igneous activity. Thermally induced stresses in the lunar lithosphere, too small to have caused rupture, can be expected to have changed from extensional, when heat generation rates were relatively high, to compressional as heat generation rates declined. Solomon and Head suggest that the cessation of rille formation records the transition from global extension

to global contraction about 3.6 b.y. ago (see Fig. 9.5.16).

The tectonic interpretation of structural features within the maria is not without controversy. As many as four different sets of ridges and arches within a lunar mascon basin have been recognized (Maxwell, 1978): (1) a circular inner-ring system; (2) ridges oriented radially to the basin; (3) north-trending ridges; (4) local systems associated with post-mare craters. Muehlberger (1974) advocated the idea that thrust faulting due to global contraction produced the ridges in southeastern Serenitatis. This is consistent with a hypothesis presented by Dvorak and Phillips (1979) suggesting that the topographic trough occupying western Mare Tranquillitatis is synclinal, resulting from global contraction of the lithosphere. Although Bryan (1973) had suggested that mare ridges formed by local sagging of the central portion of the Serenitatis Basin, Muehlberger (1977) found that only a fraction of the required crustal shortening could be accounted for by basin subsidence alone. Other evidence, however, also indicates that basin subsidence did occur during the emplacement of mare basalts. Based on the flow directions of sinuous rilles in the maria, Scott *et al.* (1978) inferred large topographic variations in the mare surfaces to represent multiple periods of tectonic deformation. Such an interpretation was corroborated by the lunar radar sounder experiment, which detected subsurface reflections in Serenitatis. The apparent subsidence of these reflectors suggests that deformation occurred prior to the last episode of basin infilling (Maxwell, 1978). Thus, for Serenitatis at least, the ridge system appears to be the result of both global stresses and local loading of the lithosphere.

Arcuate rilles peripheral to the circular lunar mare have been interpreted as grabens formed in response to subsidence of the filled lunar impact basins (e.g., Baldwin, 1963; Ronca, 1965; Quaide, 1965; Fig. 6.5.3). The existence of lunar mascon gravity anomalies (Muller and Sjogren, 1968) is consistent with the interpretation that the load of volcanic fill is wholly supported by the strength of the lithosphere (Phillips and Lambeck, 1980), but leads to local near-surface extensional failure. Several quantitative interpretations of arcuate rilles and mare ridges, based on flexure of an elastic lithosphere shell, have been reported (Melosh, 1978; Solomon and Head, 1979, 1980a). As discussed above, Solomon and Head (1979, 1980a) present an interpretation consistent with mascon loading and secular evolution of global stresses.

Melosh (1978) analyzed the tectonics of mascon loading and predicted an inner zone of radial thrust faults surrounded by an annulus of strike-slip faults

encircled by concentric normal faults. This analysis, which yields a different result from that of Solomon and Head, did not consider modifications of the stress field by faulting, pre-existing zones of weaknesses in the lithosphere, residual stresses, regional stresses or global stresses. Further, Melosh's analysis, based on the principal stress-faulting theory of Anderson (1951), fails to accommodate the possibility of extensional failure. Photogeology could perhaps be applied in the future in an attempt to discriminate between the models of Solomon and Head and of Melosh. It is worth noting that as yet only Tjia (1970) has suggested that mare ridges near the edge of mare basins involve strike-slip motion, although the difference between central mare-ridge trends and peripheral ridge trends reported by Worrall *et al.* (1978) may be tectonically significant.

The thickness of the lunar lithosphere modelled as an elastic layer over a hydrostatic interior has been estimated using evidence from the mascons. Variations in lithospheric thickness as functions of both age and location have been inferred. Solomon and Head (1980a) provide the most comprehensive analysis to date. They treated the volcanic and tectonic histories of eight major lunar mascons by considering that these histories are sensitive both to the thickness of the lunar lithosphere and to the presence of any superposed regional or global stress field. Estimates were obtained for the thickness of the elastic lithosphere in each basin region during the period of rille formation and during the later period when only mare ridges were formed. Substantial spatial and temporal variations in lunar lithospheric thickness emerged. Elastic lithospheric thickness during the interval 3.6–3.8 b.y., the time at which linear rilles formed, ranged from 25 km (at Grimaldi) through about 50 km (at Orientale, Serenitatis, Humor and Imbrium) to 75 km or more (at Nectaris, Crisium and Smythii). Younger lithospheric thickness estimates (3.0–3.2 b.y. ago), with the exception of Grimaldi, were more than 100 km.

Factors influencing the derived variation in the earlier lithospheric thickness include age and size of basins, rate of basin filling, age of basin filling and temperature variations independent of the history of basin formation and fill. The first four factors cannot among them account for all the calculated variations in lithospheric thickness, and it is necessary to infer regional heterogeneity. The western nearside appears to have had a regionally thinner elastic lithosphere at the time of mare basin formation and filling than the eastern nearside (with Nectaris, Crisium and Smythii). Perhaps significantly, this observation correlates with distributions of both volcanism and radioactivity (Whitford-

Stark and Head, 1980; Malin, 1974; Wood and Head, 1976; Trombka *et al.*, 1977).

The lunar lithosphere apparently increased in thickness over about 700 m.y. by a factor of about 2 or 3 and this indicates a general cooling of the outer part of the Moon, and of the mascon mare basins in particular. Variability in thickness appears to have decreased with time, perhaps indicating that shallow sources were responsible for much of the earlier heterogeneity.

The relative lack of mare flooding in some young multi-ringed impact basins, particularly on the farside, perhaps can be understood in the light of the heterogeneity recorded by Solomon and Head (1980a) in the thickness of the lunar lithosphere at this time. At the time of formation of these young basins, the lithosphere was probably colder and more depleted in basalt material than earlier in lunar history, and it was these depleted regions which could no longer yield basalt to flood the basin interiors. The distribution of the depleted regions was presumably established early in lunar history during a time for which the tectonic record has subsequently been obscured by the effects of the high impact flux. Alternatively, the lack of farside volcanism could be attributed to the offset, towards the farside, of the center of figure from the center of mass. If the heights reached by basaltic magmas on the lunar lithosphere were controlled hydrostatically, then the ratio of extruded to intruded basalt would be less on the farside, and extruded basalts would only reach the topographically lowest regions, as is observed.

The result of all of these tectonic studies does not totally clarify the relationships between impact-induced fractures, lithospheric stress state, and basaltic extrusion. An important result that emerges from Solomon and Head's studies is that most of the volume of the mascons is formed by rapid and voluminous basalt flooding within 0.1 to 0.2 b.y. after impact. Mare basalts are not, therefore, like basalts on Earth, recording the long-term evolution of a planetary mantle, but are predominantly erupted in response to impact and the attendant stress relief associated with the creation of a topographic low. As such they do not record a comparable history of tectonic development. The fact that these impact-induced fractures remained open to allow basalt migration to the surface attests to the global extensional regime of the lunar lithosphere. Once a fraction of the load was emplaced, the circumferential grabens may have formed and served to localize the regions of extrusion. Although the Solomon-Head model predicts a transition to a lithospheric compressional stress regime about 3.6 b.y. ago, the final 900 m of basalt fill in Serenitatis was emplaced approximately 3.2 b.y. ago (Peeples *et al.*, 1978). Thus, the mechanism of extrusion

late in lunar volcanic history, during a period of global compression, requires clarification.

6.5.2 Tectonics of basaltic volcanism on Mars

Introduction

Although the occurrence of volcanoes and volcanic rocks on Mars is abundantly clear from the images returned to Earth, there is no certainty that the rocks forming these volcanoes are basaltic in composition (see, for example, McGetchin and Smyth, 1978). For the purposes of this section, the assumption is made that they are sufficiently like basalts to be comprehended within the terms of this survey.

The intermediate place of Mars in planetary volcanism has long been clear (see, for example, Mutch *et al.*, 1976, p. 319). Lunar volcanism persisted only until 3 b.y. ago and was then mostly focused on the sites of the last great impacts. Terrestrial volcanism and that on Io persist today, on the latter with continued ubiquitous vigor and on the former, concentrated at plate boundaries. Mars records volcanic activity like that of the early Moon on part of its surface, but also shows evidence of activity persisting until relatively recently, long after major impacts ceased to occur.

The surface of Mars falls into two morphologically distinct hemispheres: plains in the north and cratered terrains in the south, separated approximately along a great circle inclined at 35° to the equator. This hemispheric asymmetry of terrain types on Mars may be a consequence of an early martian tectonic evolution that is distinct from that of the Moon and Mercury. An isostatically compensated elevation difference of about 3 km between the northern plains and southern cratered highlands leads to the hypothesis of different crustal thicknesses and/or densities in the two regions (Phillips and Saunders, 1975; Mutch *et al.*, 1976; Phillips and Ivins, 1979; Wise *et al.*, 1979a).

Isolated regions of what may be highlands terrain occur within the northern plains. Fretted terrain at the boundary between highlands and plains has been interpreted to be a result of erosional retreat of the highlands, although stratigraphic relationships do not necessarily support this view (Scott, 1978). The northern plains are generally interpreted as lava flows with varying amounts of eolian cover. Certain morphologic features may be tectonic in origin, but their interpretation is highly ambiguous (Scott, 1979).

Although it has been suggested that erosional processes produced the northern plains, it has also been argued that the elevation difference cannot be explained

simply by removing crustal material from the plains and redepositing it in the highlands (Phillips and Ivins, 1979; Wise *et al.*, 1979b). If martian crust and mantle densities were comparable to those of continental crust and mantle of the Earth, at least a 20 km thickness of material would have to be transferred from the northern plains to the highlands to account for the present elevation difference and maintain isostasy. Therefore, if Mars had an early crust of nearly uniform thickness and composition as is thought to occur on the Moon, an internal process must be strongly coupled to the formation of the asymmetry.

Alternatively, because Mars is a planet which has escaped both large-scale horizontal lithospheric motion and subduction (Phillips and Ivins, 1979), at least since the high impact influx ended (before that time there is no record), the present morphological division is one that might have persisted since early martian history (~3.8 b.y. ago). That is, thick, light, more differentiated crust apparently could have been made early over an area of about half the planet in contrast with both the Moon, which was probably entirely covered by highlands, and the Earth which is even now only one-third covered by continents. Because of the Earth's lithospheric mobility, its continental third is usually in several pieces; however, these fragments were nearly all assembled into one landmass during the short life of Pangea (~70 m.y.), ending 180 m.y. ago, and may also have been on previous occasions. On Mars, the concentration of light crust in one hemisphere may have happened stochastically. Mobility was perhaps lost before the concentration could break up.

Ancient volcanism

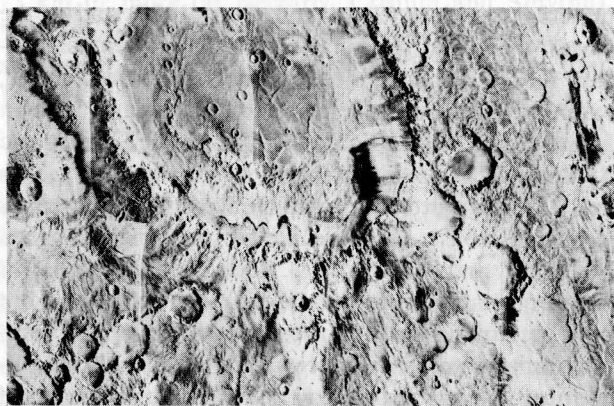
The heavily cratered terrain of the southern highlands is thought to be the oldest widely exposed surface material on Mars (Scott and Carr, 1978) and is the closest analog to the lunar highlands. Despite significant modification by the early martian cratering flux, extensive plains units believed to be volcanic in origin have been identified (Scott and Carr, 1978; Greeley and Spudis, 1978). Complex patterns of mare-like ridges and flow scarps, together with numerous volcanic cones are identifiable in the plateau plains material described by Greeley and Spudis (1978). The ridges in these plains lack any recognizable pattern in their distribution, but may perhaps indicate that regional (or global?) compressional tectonism was prevalent during early modification of the martian highlands. Numerous examples of floor-fractured craters have been identified by Schultz and Glicken (1979), and their distribution correlates well with the location of the highland-lowland boundary. As on the Moon, floor-fractured craters are inter-

puted to be examples of volcanically modified impact craters and, from their varied degradational states, indicate that highland volcanism was a protracted process on Mars.

Martian basin volcanism

In addition to the plains-forming units within the southern highlands, several old volcanic constructs (patera and tholii) have been identified. Australis Tholus (unit 47 in Chapter 8) and Tyrrhena Patera (unit 45) are both surrounded and embayed by floor volcanics that display complex distributions of wrinkle ridges. Peterson (1978), in discussing the Noachis-Hellas volcanoes, suggested that the occurrence of five volcanic centers in that region may be localized on the second and third basin rings of the Hellas structure where the rings are intersected by a northeast-trending deep fracture system of which there is some evidence in the gravity field. According to Peterson, localization of activity by the rings might imply that volcanism started not long after basin formation, but relatively youthful-looking features indicate either persistent or recurrent volcanism around Hellas. Radial grabens may also exist around the basin Schiaparelli (Fig. 6.5.4) and are genetically similar to features around Mare Imbrium on the Moon.

More recognizable as basin-related tectonic features are the concentric ridge systems within Schiaparelli and Huygens basins. These features may result from subsidence during basin loading similar to the mode of formation of lunar mare ridge systems (Solomon and Head, 1980a). Circumferential grabens, similar to those around Mare Serenitatis, are also present around Isidis basin on Mars, which contains a lunar-like mascon (Sjogren, 1979). The gravity field of Isidis has been



6.5.4 Mare ridges and possible radial grabens characterize the martian basin Schiaparelli. Viking Orbiter photo-mosaic.

analyzed as a flexural load and indicates an elastic lithosphere thickness formation of about 100 km (Solomon *et al.*, 1979). Although the volcanic origin of martian basin fill remains unproven, by analogy with lunar tectonic features, it appears that the martian basins may also mark the location of voluminous magma eruptions at an unspecified time after basin formation.

Tharsis province: volcanism

Later martian volcanism is generally confined to two broad regions: Tharsis and Elysium. The volcanic structures of the Elysium area are generally less degraded than those of Hellas but not as fresh as those of the Tharsis province. The Tharsis province, because of its size, constitutes one of the most tectonically remarkable volcanic structures in the solar system.

Elysium and Tharsis both form large-scale topographic (but not necessarily structural) domes that rise 6 and 10 km above the surrounding plains and have diameters of 40° and 90° of the surface arc, respectively. The Tharsis Dome (about 4000 km across) is flanked by depressions in Chryse and Amazonis Planitiae. Both domes have associated positive free air gravity anomalies and the flanking depressions of Tharsis have negative free air anomalies. The Tharsis anomaly dominates the second and third degree harmonics in the global gravity field (Phillips and Saunders, 1975; Lambeck, 1979; Sjogren, 1979), thus strongly emphasizing its global importance.

Both the Tharsis and Elysium provinces are capped by several individual volcanoes that rise in height an additional 5-15 km to a maximum elevation of approximately 25-27 km above the mean Mars datum. Multiple periods of caldera collapse, producing summit craters as much as 100 km in diameter, are evident on these volcanoes. Many individual lava flows on the flanks are in excess of 500 km in length (for an extended review of martian volcano morphology, see Chapter 5).

Tharsis province: interior structure

Evidence of the relative ages of late martian volcanism comes from cratering intensity studies as well as from structural and volcanic stratigraphic relationships. Although we have no reliable way of tying martian crater chronology to that of the isotopic age determinations on the Moon, we can probably assume that the high impact flux died away at about the same time for both objects. From this it appears that the later structural and volcanic events of Mars extended over a lengthy period (see Chapter 8). This implies that the total stress transmitted to the lithosphere at any one time was probably readily withstood without shear failure.

As there are no plate boundaries on Mars and the lithosphere is in one piece, later martian volcanoes are most like large intraplate volcanoes on Earth, such as those of Hawaii and Northern Africa. The Hawaiian and African volcanoes may reflect underlying mantle heterogeneities or may be responses to lithospheric fracturing; both hypotheses have their supporters.

Geophysical data, mainly gravity and topography, and surface structural features provide important constraints on the genesis and present interior structure of Tharsis. As reviewed by Mutch *et al.* (1976), Tharsis has been explained as an updomed region of the surface. This model is supported by the high elevation of old, densely cratered surfaces.

The radial pattern of Tharsis faulting (see below) has also been cited as evidence for doming. Wise *et al.* (1979a) and Plescia and Saunders (1979b), however, suggested that these features represent superimposed linear trends rather than a contemporaneous radial pattern. Alternatively, the fracture system may be a response to loading.

The relative importance of updoming in the Tharsis region has recently been questioned by Solomon and Head (1980b). They pointed out that the traditional view that the elevated, densely cratered terrain of Tharsis implies updoming of an earlier surface formed at lower elevations is not required by photogeologic observations. It is proposed that Tharsis and Elysium may be equally well explained as thick constructional volcanic piles partially supported by an elastic lithosphere.

Early studies of gravity and topography indicated that the topography could be only partially supported by shallow compensation due to variations in crustal thickness or density (Phillips and Saunders, 1975). Possible dynamic support of the Tharsis region by mantle convection is discussed by Phillips and Ivins (1979). An axially symmetric, first harmonic mode of thermal convection in the sub-lithospheric mantle with upwelling beneath Tharsis would be suggested. The absence of an antipodal, basin-like feature with a negative gravity anomaly might be explained by the diffuse nature of the downgoing convective flow that results from the temperature dependence of viscosity. Another mechanism for localizing convection beneath Tharsis is a concentration of heat release due to core formation. Wise *et al.* (1979a) speculated that a mechanism involving mantle convection and core formation resulted in subcrustal erosion beneath the northern plains and the emplacement of this crustal material beneath Tharsis.

Sleep and Phillips (1979) have shown that an isostatic model will satisfy the gravity data if the Tharsis province can be described in terms of a thin crust and a low-density upper mantle. Typical values, with a litho-

spheric thickness greater than 300 km, are a martian crust about 100 km thick, in general, thinning to about 50 km in the center of Tharsis, and an upper mantle (above about 300 km) beneath Tharsis contrasting in density with normal martian upper mantle by about 0.2 gm/cm³. Sleep and Phillips (1979) have suggested that there is a resemblance between Tharsis volcanism, elevation and gravity field and those of the Basin and Range province. As also pointed out by Sleep and Phillips, a much closer resemblance exists to the great uplifts of Northern Africa described by Burke and Whiteman (1973) and, following them, Fairhead (1979), who suggested that these uplifts were precursors of tectonics of the African Rift valley type.

The terrestrial analogues to Tharsis are sites of active volcanism. The uplifts are dynamically maintained because of the underlying partial melting, and the lithosphere is thin under both the Basin and Range province and the African uplifts. Fairhead and Reeves (1977) inferred this for the African areas by correlating P delays for Kazakhstan nuclear tests with Bouguer anomalies; Burke and Whiteman inferred it simply because the erupted basalt comes from depths of about 50 km. If, in the future, volcanic activity ceases on the African uplifts, the elevations will be rapidly removed by erosion, the underlying mantle will revert to being part of the lithosphere (consequently becoming denser) and the associated gravity anomaly will not persist. In the Sleep and Phillips (1979) model, martian lithosphere under Tharsis is regarded as rather thick to maintain lateral density inhomogeneities, and the terrestrial analogues cannot be applied. However, dynamically maintained, low-density upper mantle is also consistent with the model, but if the terrestrial analogues are to be pressed very far, evidence of currently active volcanism at Tharsis seems necessary.

Tharsis province: lithospheric evolution

It seems probable that the thickness of the martian lithosphere in the Tharsis area has varied with time. By analogy with Africa and the Basin and Range, the lithosphere in Tharsis thinned during the time of volcanic activity. Locally thinned lithosphere causes stress concentration and localizes rupture, but it is not usually possible to tell whether the rupture happens first and volcanism is concentrated at the fractures, or whether magmatism first thins the lithosphere. In Tharsis there is some evidence of faulting earlier than much of the volcanism. Whatever the sequence of events, a locally developed, thin lithosphere would have characterized Tharsis during active volcanism and the huge load would have been maintained by the underlying, mass deficient, partly melted mantle. When volcanic activity

ceased at Tharsis, the lithosphere probably reverted to the contemporary thickness of the lithosphere over the rest of the planet. Consequently, if the elastic lithosphere below Tharsis is supporting the vast load without continuing extensive failure (and this seems to be the case), then it may prove to be some hundreds of kilometers thick.

Other estimates for martian lithospheric thickness may be obtained by assuming that the magma source region for surface volcanism was located at the bottom of the lithosphere. The hydrostatic rise of magma to the summit of Olympus Mons suggests a magma source region at a depth of only a few hundred kilometers (Vogt, 1974a; Phillips and Ivins, 1979). Several studies (Carr, 1974; Vogt, 1974b; Plescia and Saunders, 1979a) have indicated increasing height and spacing of Tharsis volcanoes with decreasing inferred age. Plescia and Saunders (1979a) in particular have noted that older shield volcanoes were small low features while younger shields are significantly broader and taller. Following the earlier suggestion of Vogt (1974a) for terrestrial oceanic islands, it is interesting to speculate that increasing volcano heights with decreasing relative age are a result of a lithosphere that thickens with time.

Tharsis province: tectonic deformation

Four kinds of evidence are relevant in considering tectonic character of the later volcanism on Mars: the associated structural style, the gravity field, the distribution of the volcanism and its duration.

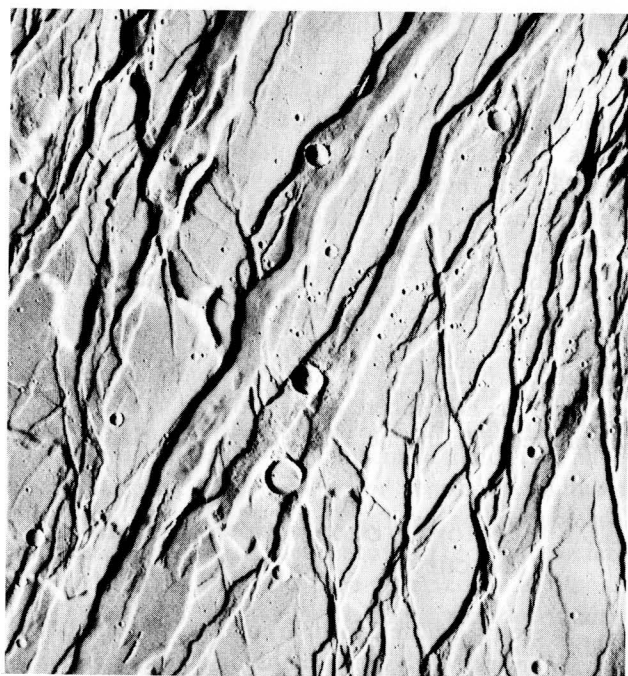
The surface of Mars is highly fractured (Mutch *et al.*, 1976, p. 226) and very many of these fractures are associated with extension and with graben formation. This structural pattern is best known in the Tharsis plateau where radial swarms of grabens extend south, southwest, north, and northeast of the uplift for as much as 2500 km (Fig. 6.5.5). Such grabens are 2-5 km in width, and in places (e.g., Tempe Fossae; Fig. 6.5.6) form a broad belt of fractures about 500 km wide. There appears to have been a long history of deformation in the plateau region where several sets of grabens show different orientations. Wise (1974b) documented a sequence of events in the Alba and Tempe regions on the northeastern fringe of the Tharsis uplift that involved five geologic episodes including two volcanic episodes separating and succeeded by fracturing and faulting.

Complexes of interlocking grabens are particularly well-developed in the Noctis Labyrinthus area. The most prominent fracture system radiating from Tharsis is the huge Valles Marineris whose straight walls attest to a structural origin; it has been compared to the great rifts of Africa (Hartmann, 1973). In addition, an extensive series of mare-like wrinkle ridges is located to the east of the Tharsis dome in Lunae Planum (Fig. 6.5.7). Those ridges have a pronounced north-south orientation, are typically 300-400 km in length, and have a regular spacing of approximately 400 km.

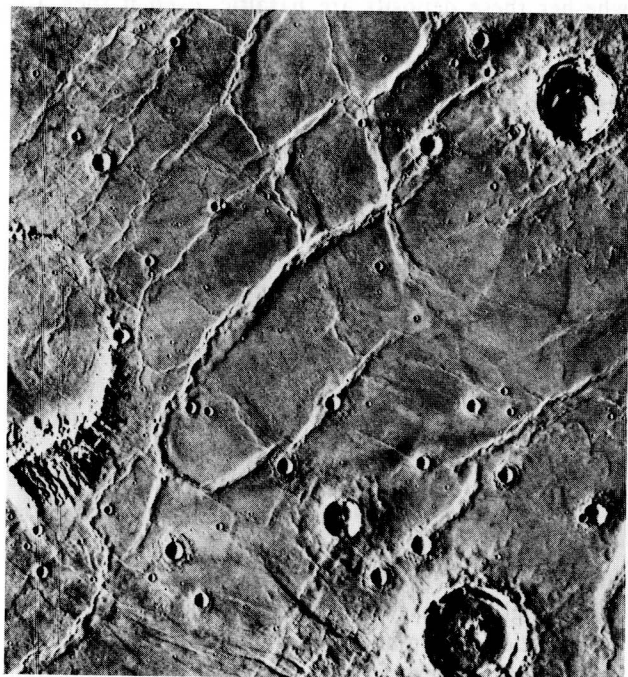
Several recent studies examine the development of the Tharsis tectonic features. Wise *et al.* (1979a) identified two main centers of faulting: (1) an early center in the Thaumasia area southeast of the Tharsis dome with four distinct fault systems with linear trends and (2) a younger, northwest-southeast, elongated, radial set with a center near the crest of the present Tharsis dome. Lunae Planum wrinkle ridges are reported to form an elliptical pattern about a different center. Based primarily on features in the vicinity of Valles Marineris, Frey (1979a) also identified a dome in Thaumasia as an early center of faulting and uplift. He described the faulting as radial about the Thaumasia and Tharsis domes, and suggested that the Lunae Planum wrinkle ridges are radial to the Thaumasia center. Plescia and Saunders (1979a) also believed that they could identify two centers of fracturing, one near Pavonis Mons and the second within Syria Planum.

Despite the lack of general agreement on the history of faulting, it is clear that several periods of faulting generally radial to Tharsis took place. Apparently, the largest of these faulting episodes produced the grabens in Memnonia and Tempe Fossae with the possible initiation of the Valles Marineris rifting (Wise *et al.*, 1979a; Frey, 1979a). In addition, the three Tharsis summit volcanoes (Arsia, Pavonis and Ascræus Montes) all lie on the same great circle, which also passes through the heavily faulted Tempe Fossae region over 2000 km to the northeast (Fig. 6.5.8). This suggests that fractures may have provided an important control on surface volcanism.

The tectonic development of Tharsis has been compared to the East African Rift System and to the tectonic province which comprised portions of North America, South America and Africa during the initial



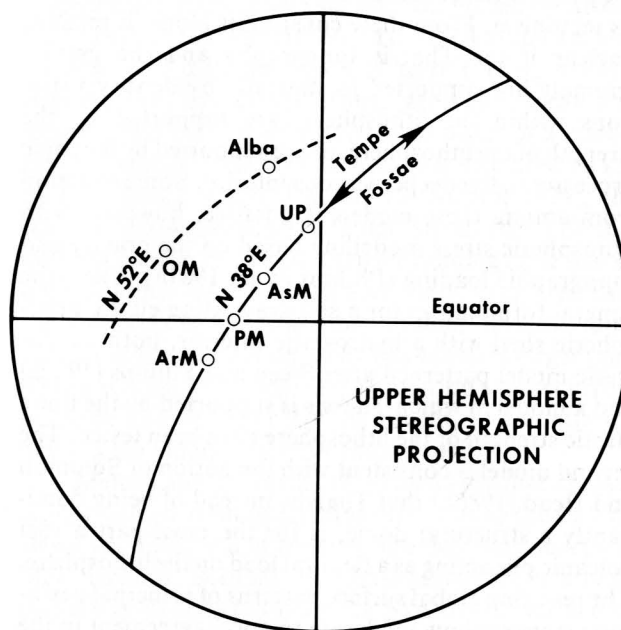
6.5.6 Multiple grabens 2000 km northeast of Tharsis are typical of the Tempe Fossae region on Mars. Viking frame number 627A36.



6.5.7 Wrinkle ridges morphologically similar to lunar mare ridges are common features in the Lunae Planum region of Mars. Viking frame number 610A22.

opening of the Atlantic Ocean (see Mutch *et al.*, 1976). Possible similarities have been identified between the Valles Marineris rifting episode and features observed in the East African Rift (Frey, 1979b). Several attempts have been made to correlate the creation of the Valles Marineris (Coprates) canyon system with incipient plate tectonics on Mars (Courtilot *et al.*, 1975; Frey, 1979a; Schonfeld, 1979). This idea of incipient lithospheric spreading has, however, been dismissed by Sharp (1973), Phillips and Ivins (1979), Carr (1980) and Frey (1979a).

Nearly all structures recognized around Tharsis are extensional. Şengör and Jones (1975) and a very few other workers have drawn attention to sigmoidal structures which are strongly suggestive of horizontal shear in the Tharsis region and they, almost alone, have also seen evidence of convergent structures. The wrinkle ridges, seen in Lunae Planum, are probably compressional in origin, but are minor features when contrasted with the extensive graben systems. Operation of planet-wide tectonics with a predominance of extensional



6.5.8 Alignment of the Tharsis Montes line of volcanoes. Locations of major volcanoes have been plotted on the upper hemisphere of a stereographic projection with 0° longitude at the right and 180° longitude at the left. The four volcanoes lie within a half-degree of a great circle oriented N38E. Tempe Fossae lies on the northeast projection of the same great circle. Olympus Mons and Alba Patera lie on a different great circle. Abbreviations: OM, Olympus Mons; UP, Uranus Patera; AsM, Asraeus Mons; PM, Pavonis Mons; ArM, Arsia Mons. From Wise *et al.* (1979a).

features is possible because rocks are stronger in compression than in extension.

A variety of alternative interpretations based on Tharsis tectonic features has been proposed. Solomon and Chaiken (1976) emphasized the possible relationship between extensional tectonics and volume change due to planetary thermal evolution. Mutch *et al.* (1976) considered that expansion might be due to phase transitions in the mantle. McAdoo and Burns (1975) suggested that polar wandering of the lithosphere might also explain tensional faulting in the Valles Marineris complex. Melosh (1980) has also recently discussed tectonic patterns that might have resulted from such a reorientation of the lithosphere relative to the rotation axis. The movement of the excess mass of Tharsis toward the rotational equator of the planet, as a reorientation of the principal moment axes, is a possible driving mechanism. These studies, which use a faulting criterion different from that of McAdoo and Burns (1975), suggest that extensional structures radial to Tharsis are not related to polar wandering.

Gravity and topography only partially constrain the possible range of models which might explain Tharsis tectonism. From these constraints alone, it remains unclear if the Tharsis topography and the gravity anomaly are supported isostatically by density variations within the lithosphere, are supported by the strength of the lithosphere, or are supported by dynamic processes in the deeper viscous interior. Some selection from among these models is possible, however, with lithospheric stress modelling based on the gravity and topographic loading (Phillips *et al.*, 1981a). Using the general formulation for a self-gravitating elastic lithospheric shell with a hydrostatic interior, both an isostatic model patterned after Sleep and Phillips (1979a) and a model in which Tharsis is supported by the finite elastic strength of the lithosphere have been tested. The second model is consistent with the notion of Solomon and Head (1980b) that Tharsis, instead of being dominantly a structural dome, is for the most part a vast volcanic pile acting as a flexural load on the lithosphere. The resulting global surface patterns of principal deviatoric stresses show moderate to good agreement in the orthogonality of tensile stresses with inferred tensional surface features (e.g., grabens) and of compressive stresses with inferred compressional features (e.g., wrinkle ridges in Lunae Planum). However, the flexural model is clearly superior in comparison to the isostatic model in its agreement with the geologic features, indicating the likelihood of at least partial support by the finite strength of the lithosphere.

Whichever model (flexural or isostatic) is correct, however, a key result is that the tectonic features asso-

ciated with Tharsis, including Valles Marineris, are clearly related to the incumbent loading of the lithosphere, and are not the result of structural updoming, which would have produced the conjugate set of structures. The lack of angular isotropy and contemporaneity of the fractures most likely reflects lithospheric heterogeneity (and evolution thereof). While the preferred (flexural) model is suggestive of volcanic constructional loading, significant internal loading (intrusives, mantle thickening, etc.) is not ruled out. Independent of its origin, Tharsis topography, once created, would tend to sustain access of magmas through the lithosphere because of the tensile stresses associated with the mere presence of the relief itself.

6.5.3 Tectonics of basaltic volcanism on Mercury

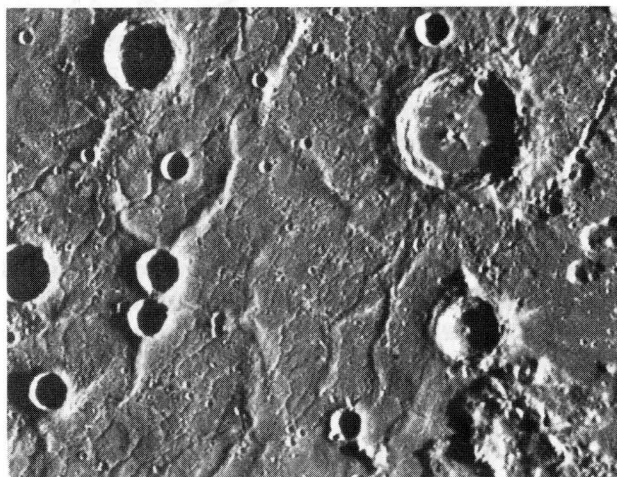
Mercury is little known in comparison with Earth, the Moon and Mars, so consideration of tectonics in relation to volcanism is extremely speculative. Mariner 10 images provide the sole basis for studies of the mercurian surface. For example: Mercury's smooth plains may be formed either of volcanic material (Trask and Strom, 1976) or of ejecta (Wilhelms, 1976; Oberbeck *et al.*, 1977), and it is clearly very difficult to know whether these deposits are basaltic, even if they are volcanic (Malin, 1978).

Strom *et al.* (1975a) suggested that irregular scarps restricted to crater floors may be lava flow fronts. If so, their heights of 100–500 m show that they were formed by very viscous lava compared to that of the lunar maria. A strong argument in favor of volcanic activity at some time in Mercury's early history is that the planet's magnetic field requires at least a partially molten metallic core. Formation of such a core might dissipate enough energy to ensure volcanic activity.

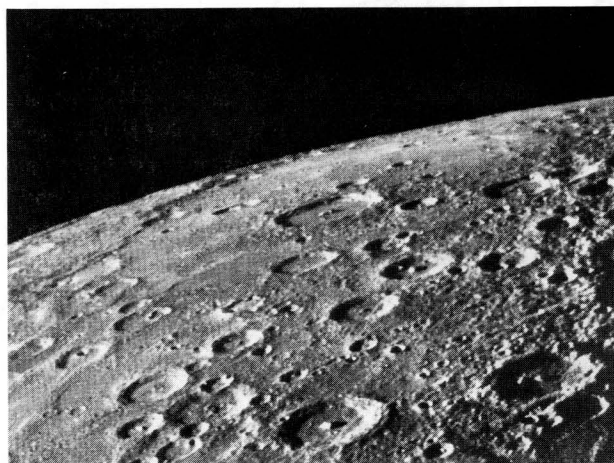
The absence of recognizable surface features which can be related to plate tectonics has led to the general consensus that Mercury, like the Moon, has had a mechanically continuous lithospheric shell dating at least from the end of the period of heavy bombardment. Tectonic features on Mercury show some similarities but also important differences from those on the Moon. Unlike those of the Moon, the major tectonic features are thought to have developed in response to tidal deformation and planetary volume change as well as basin loading (Burns, 1976; Melosh, 1977; Pechmann and Melosh, 1979).

If there is volcanic activity recorded on the surface of Mercury, the passage for the magma to the surface is probably, like that on the Moon, overwhelmingly

linked to fracturing associated with impact cratering. There are also resemblances in morphology and distribution between the smooth plains of Mercury and the lunar maria (Strom *et al.*, 1975a). One very large crater, 1300 km in diameter, the Caloris Basin, was surveyed by Mariner 10. Wrinkle ridges similar in morphology to those on the lunar maria are developed in smooth plains units within this basin (Fig. 6.5.9). These ridges have orientations radial and concentric to the basin and are interpreted as compressional features (Strom *et al.*, 1975a). Ridges range in width from 1 to 12 km, attain heights of 500 m, and are continuous along their strike for as much as 300 km. Within the Caloris Basin, the ridges are cut by trough-like features forming a polygonal pattern with radial and concentric trends that have been interpreted as grabens or tensional fractures (Strom *et al.*, 1975a; Fig. 6.5.9). The fractures become progressively wider and deeper toward the basin center. Similar features have not been observed on the lunar maria. The wrinkle ridges and troughs in smooth plains within and around the Caloris Basin have been interpreted as structural features developed in response to basin tectonics (Maxwell and Gifford, 1980). Wrinkle ridges are hypothesized to be compressional features formed due to subsidence of the basin resulting from the loading of smooth plains-forming volcanics. Fractures or grabens, with trough-like morphologies, are related to uplift of the central portion of the basin. These fractures may have provided the only conduits for basaltic magmas to reach the surface.



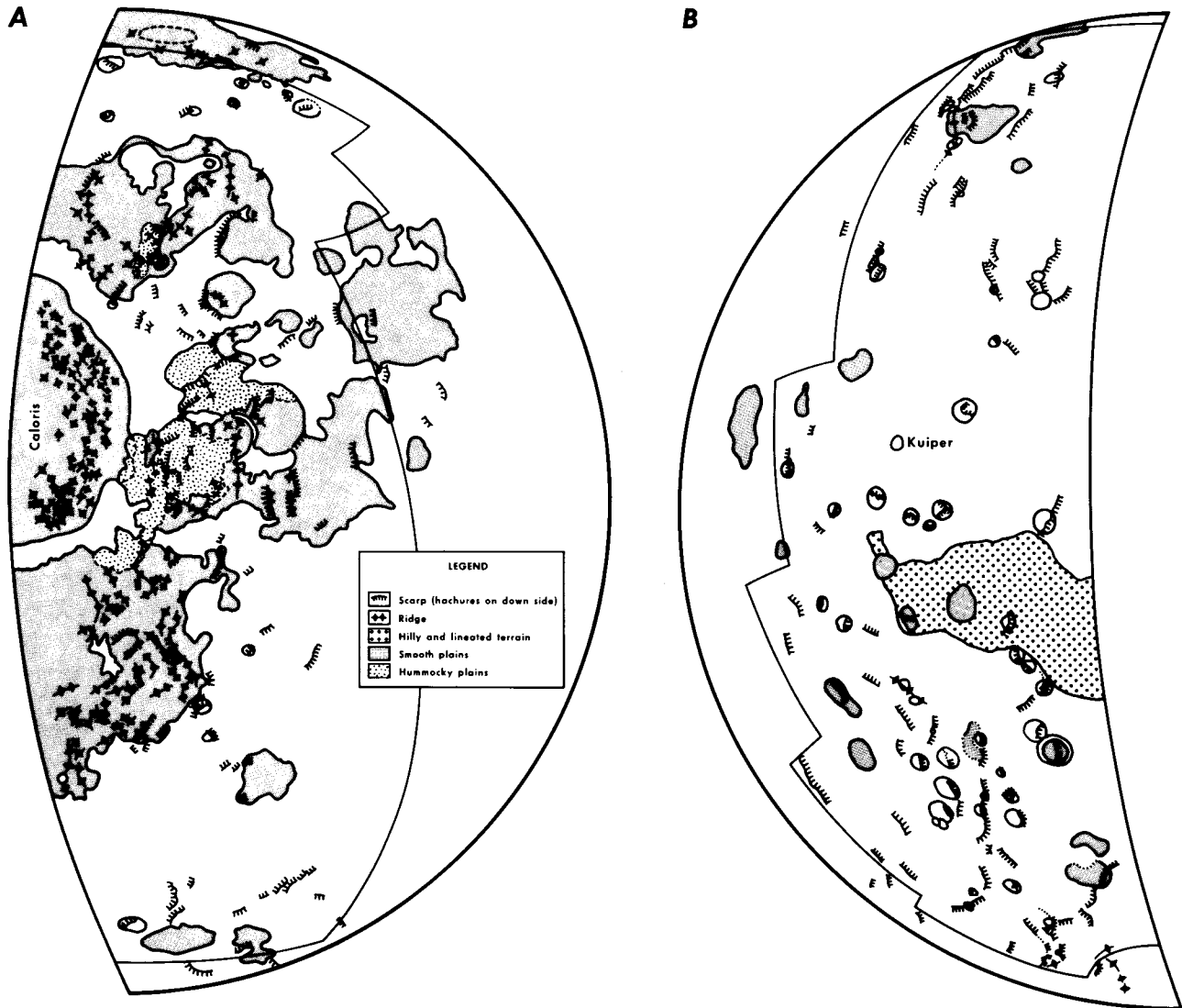
6.5.9 Part of the floor of the Caloris Basin showing the ridges and fractures. The length and width of the fractures increase toward the center of the basin (lower right to upper left) and the fractures transect the ridges at various angles (photo FDS 126). After Strom *et al.* (1975a).



6.5.10 Examples of scarp on Mercury extend off to the horizon in this Mariner 10 image (photo FDS 27328).

The 30% of Mercury imaged by Mariner 10 does, unlike the Moon, show evidence of a regional, perhaps even global, tectonic development (Fig. 6.5.10). This evidence lies in the lobate scarp, recognized by Strom *et al.* (1975a), that vary in length from 20 to 500 km and cross several different types of terrain (Fig. 6.5.11). Because they are cross-cutting, generally linear and lengthy, and because of a horizontal offset of a crater rim (the wall of crater Guido Arezzo is displaced 10 km by the Vostok scarp, Fig. 6.5.12; Strom, 1979), the scarp are considered to be faults. Strom *et al.* (1975a), noting a contrast between scarp morphology and that of normal faults on the Moon and Mars, suggested that the scarp on Mercury are thrust faults. Because some of these scarp are disrupted by fresh craters larger than 40 km, it is believed that scarp formation was occurring near the end of heavy bombardment, and continued beyond the emplacement of smooth plains.

Melosh and Dzurisin (1978) indicated that arcuate scarp formed in smooth plains surrounding the Caloris Basin are smaller and fewer in number than scarp in older terrains. Because the lobate scarp are relatively uniformly distributed over the part of the surface of Mercury viewed by Mariner 10, it seems possible that an overall decrease in the surface area of the planet may be recorded in this thrusting. Strom *et al.* (1975a) estimated a decrease equivalent to a 1–2 km reduction in the radius of Mercury. Solomon (1976) calculated a comparable decrease by considering the thermal history of the planet (see Chapter 9). It is interesting to consider that a process which was widely considered during the 19th century to have been responsible for mountain building on Earth, but which was abandoned after the



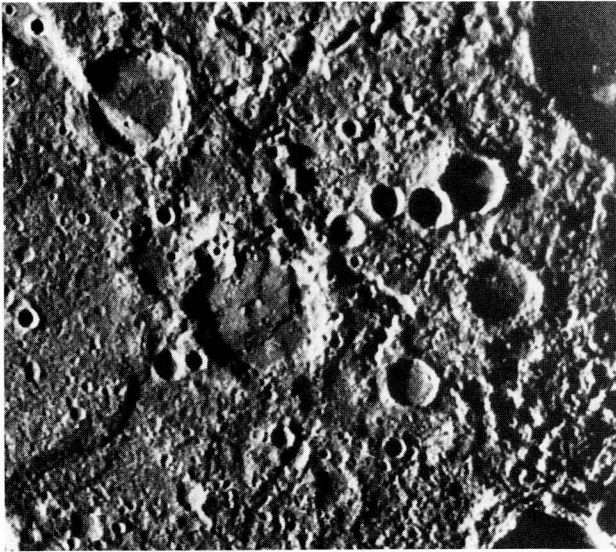
6.5.11 A tectonic map of Mercury showing lobate scarps and ridges on the incoming (b) and outgoing (a) side of Mercury. Quasi-circular features are craters. Modified from Strom *et al.* (1975a).

discovery of radioactivity, may have been dominant in tectonic evolution on Mercury.

Burns (1976) first examined the stresses and patterns of deformation that might result from tidal despinning. The combined effects of planetary volume change and tidal despinning were treated by Melosh (1977) and by Pechmann and Melosh (1979). For the tidal despinning of a planet with a thin elastic lithosphere, an equatorial zone of strike-slip faults and a polar region of east-west striking normal faults are predicted. Simultaneous contraction suppresses the region of normal faulting and introduces an equatorial zone of north-south striking thrust faults. In the absence of

stresses due to tidal despinning, contraction should result in isotropic compressional stresses within the lithosphere and in a random orientation of thrust faults on the surface.

The relative importance and timing of tidal despinning and contraction to the development of surface structural patterns remains a subject of discussion. Cordell and Strom (1977) considered that the orientation of lobate scarps is random and not the north-south orientation expected for tidal despinning. Melosh and Dzurisin (1978) suggested that the more restricted class of arcuate scarps has a preference for north-south orientation and that the northeast and northwest trending



6.5.12 Two craters (unnamed and Guido d'Arezzo) are cut by the Vostok scarp (about 130 km long). The northwestern rim of Guido d'Arezzo, 65 km in diameter, has been offset about 10 km by the scarp. This offset appears to have been caused by shortening of the crater due to thrusting of the eastern part of the crater over the western part (photo FDS 27380).

structural lineaments were produced by strike-slip faulting. They proposed that continued contraction after tidal despinning could result in the observed range of arcuate scarp orientations. Normal faulting in polar regions appears to be a critical test of the importance of tidal despinning in controlling planetary tectonic patterns. Such features have not yet been observed. Strom *et al.* (1975b) indicated that numerous lobate scarps occur in south polar regions, but the ages of these scarps relative to tectonic features elsewhere have not been established.

The fundamental lesson to be learned from photo-geologic examination of the surface of Mercury is that the lack of clearly identifiable volcanic flows on the surface of the planet is probably related to a lithosphere which has been in compression over much of the history of the planet (Solomon, 1978). Since scarp formation was occurring near the end of the period of heavy bombardment (~4 b.y. ago?), the lithosphere was probably in compression at that time or earlier and conditions were not favorable for magma migration to the surface. Very early surface volcanism, during a period of lithosphere extension, would be obscured by the high impact flux. Subsequent volcanism, if it occurred at all, must have been closely related in space and time to major impact events.

6.5.4 Tectonics of basaltic volcanism on Venus

Venusian tectonics remain less well understood than the tectonics of the other terrestrial planets due to the lack of detailed surface observations. Both Earth-based radar (Malin and Saunders, 1977) and the Pioneer Venus orbiter (Pettengill *et al.*, 1979, 1980; Masursky *et al.*, 1980a, b), however, provide low resolution topographic data. Precision tracking of the Pioneer spacecraft has also been used to determine the global gravity field (Phillips *et al.*, 1979b; Sjogren *et al.*, 1980; Ananda *et al.*, 1980).

In early Earth-based radar observations (Saunders and Malin, 1975), circular features with the gross morphologies of impact craters, one large trough-like depression comparable to Valles Marineris or to the East African Rift, and several mountainous regions of high relief were identified. Confirmation of these earlier observations and more complete coverage of surface topography has been obtained using the orbiting Pioneer Venus spacecraft (Pettengill *et al.*, 1979, 1980). From these data, Masursky *et al.* (1980) reported that about 27% of the surface of the planet consists of lowlands with elevations (relative to the mean radius of the planet) less than 0 km, 65% consists of upland plains with elevations of 0–2 km, and the remaining 8% is highlands with regional elevations in excess of 2 km. The highlands are dominated by two continent-sized regions, Aphrodite and Ishtar, with mountainous features reaching elevations of slightly more than 11 km. The uplands terrain is characterized by shallow circular features which may be impact craters and by radar-bright linear features which may be fault zones. The distribution of relief is reported to be unimodal with large areas of the surface at a nearly uniform elevation. Pettengill *et al.* (1980) indicate that the center of mass-center of figure offset is less than 400 m.

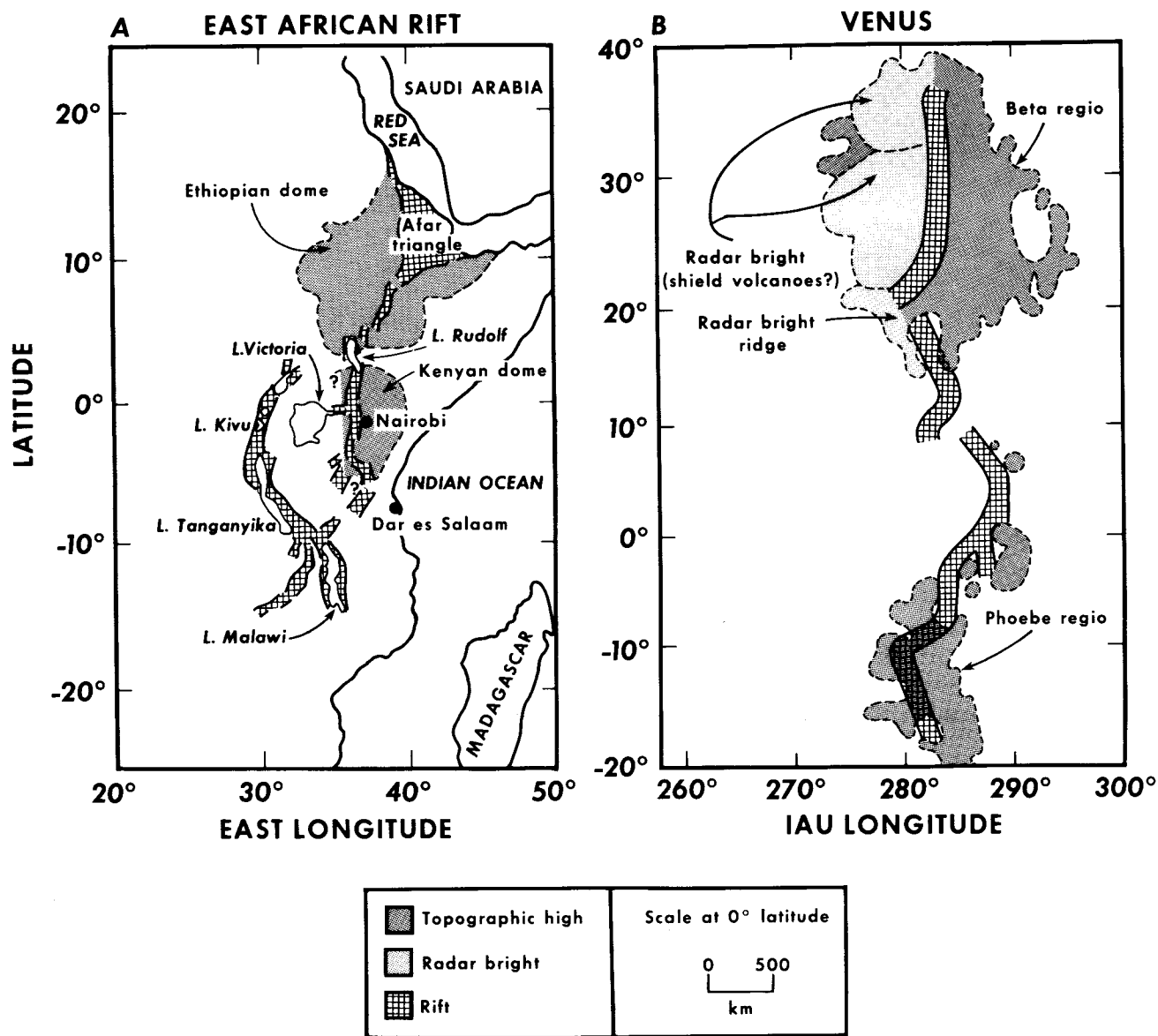
Although smaller than Aphrodite or Ishtar, the elevated Beta Regio appears to be associated with basaltic volcanism. On the basis of radar interpretation, the elevated southwestern crest of Beta may be a shield volcano (Saunders and Malin, 1977). The Venera 9 and 10 spacecraft landers determined the surface rocks on the flanks at Beta to have K, U, and Th contents similar to those of terrestrial basalt (Surkov *et al.*, 1976). Whether other highland regions are basaltic or not is problematical. Aphrodite occurs along a linear equatorial zone of elevated topography that includes Beta Regio and by inference might be associated with basaltic volcanism.

A depression to the east of the Beta "shield" has been mapped by Pioneer Venus and appears to be part

of a chain of depressions defining a trough extending southward from 35°N the full north-south length of Beta Regio (McGill *et al.*, 1981). A similar trough south of Beta is seen in Earth-based radar images (Goldstein *et al.*, 1976), so that there exists a moderately well-defined north-south elevated trend with a median trough to 20°S, and possibly as far as 45°S. When compared with the East African Rift system, the morphological similarity is striking (Fig. 6.5.13). However, the latter tectonic system is not characterized by basaltic

shield volcanism. Other “ridge and trench” and lineament systems occur on Venus (Masursky *et al.*, 1980). In total, there is ample evidence for horizontal stress in the lithosphere of Venus, but there is no way to clearly decide whether the origin of this stress is due to lithospheric plate motion or to regional loading of the lithosphere.

Gravity anomalies on Venus are relatively small compared to those for the Moon or Mars and have magnitudes more comparable to those of the Earth



6.5.13 Comparison of East African Rift System with proposed rift system associated with Beta Regio and Phoebe Regio regions of Venus. Topographically high areas that are bright in Earth-based radar images are noted; they have been proposed as shield volcanoes. (Figure from George McGill, pers. comm., 1981.)

(Phillips *et al.*, 1979; Sjogren *et al.*, 1980). In contrast to the Earth, however, there appears to be a good correlation between gravity and topography at long wavelengths. Gravity anomalies are a factor of 4 or 5 smaller than would be predicted for uncompensated topography with a crustal density of 3 gm/cm³ and are most consistent with compensation at a depth of about 100 km (Phillips *et al.*, 1981b). If Venus contains heat sources similar to the Earth, then the high Venus surface temperature implies an elastic lithosphere thickness no greater than 50 km, for even the most creep-resistant (as measured in the laboratory) geologic material. Thus the compensation of 100 km depth suggests that the topography is dynamically maintained (e.g., by convection) or is very young geologically.

If the impact crater interpretation of the upland circular features is correct, then vast areas of Venus are geologically old, and it is unlikely that present-day crustal generation by plate tectonics similar to that on Earth is taking place. An apparent absence of the characteristic concave cross-sectional topographic profiles associated on Earth with cooling diverging oceanic plate boundaries is further evidence against a present-day global system of tectonic plate motions on Venus. The linear zones of elevated topography found on Venus may be in essence the "pile-up" of new (basaltic?) crust in a lithospheric regime which does not allow spreading. An inability of the Venus lithosphere to sub-

duct crustal materials might be attributed to the present surface temperature, which is about 450° higher than that of the Earth (Phillips *et al.*, 1981b). The lithosphere on Venus might be of relatively lower density and thus more buoyant and difficult to subduct because of a deeper basalt-eclogite transition (Anderson, 1980) and because of, if all other factors are equal, a higher Mg/Fe ratio in basaltic magmas.

The lack of contemporary plate tectonics does not rule out the possibility of significant horizontal stress in the venusian lithosphere. The great elevations of the highlands regions, particularly at Ishtar, impart stresses to the interior of at least several kilobars, even if they are in isostatic equilibrium. Such great elevations might require support both from density compensation and horizontal lithospheric stress, as has been suggested for the Tibetan Plateau on Earth (Molnar and Tapponnier, 1978). Phillips *et al.* (1981b) suggested that Ishtar might be the only true terrestrial-style continent on Venus and that it might be supported by the stresses associated with basaltic loading of the lithosphere in the equatorial highlands region, made up in part by Beta Regio and Aphrodite. Thus, a variety of important factors which may govern the tectonics on Venus has been proposed. It is clear, however, that better resolution of surface features is required to provide further constraints on venusian tectonics and its relationship to basaltic volcanism.

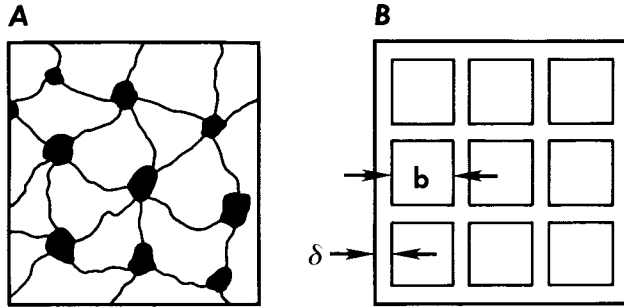
6.6 TECTONICS OF BASALTIC MAGMA MIGRATION

The origin of the magma that flows from most volcanoes on Earth is now reasonably well understood. A large fraction of the magma is produced by partial melting in the mantle, usually in the asthenosphere. (See section 1.4.1 for processes that influence basaltic magma composition. Also see section 9.4.3 for rheology.) The volcanism at ocean ridges is associated with pressure release melting during ascending mantle convection. Ascending convection of mantle rock is required to provide the material that makes up the spreading lithospheric plates. As the mantle rock ascends adiabatically and crosses its solidus, partial melting occurs. The magma then ascends through the mantle rock to form the oceanic crust.

The volcanism that forms the island arcs behind ocean trenches must be related to the descending lithospheric plate. Frictional heating occurs on the slip zone between the descending lithosphere and the overlying mantle. This heating may directly melt the descending oceanic crust including entrained sediments or it may

result in the dehydration of hydrated minerals in the downgoing slab of crust. The resulting water may flux and cause partial melting of overlying mantle rocks. If these processes occur beneath continental crust, secondary melting may occur when the ascending magmas reach the lower crust. Intraplate volcanism is less well understood. Some fraction of this volcanism may be the result of pressure-release melting in mantle plumes. An alternative explanation is that the magma reaches the surface from the asthenosphere due to lithospheric fractures under tensional stresses.

Whatever the cause of the partial melting of the mantle rock, there is strong observational evidence that the initial melt fraction will form along the intersection of grain boundaries (Waff and Bulau, 1977, 1979). This is illustrated in Fig. 6.6.1a. When the degree of partial melting has reached a few percent, the magma which has collected along grain boundaries will form interconnected channels. The magma will then be free to migrate along these channels.



6.6.1 Magma migration: **a.** Initial melt forms along the intersection of grain boundaries in a rock. **b.** A model is used to allow permeability to be related to porosity. The rock is modelled as a cubical matrix of tubes of circular section. The matrix has dimensions b and the tubes have diameters δ .

In order to model the flow of magma along grain boundaries, the magma may be considered to be a fluid in a porous matrix, the solid crystals making up the matrix. This type of model has been proposed by Frank (1968) and by Turcotte and Ahern (1978). The volume fraction of magma present defines the porosity, X , of the porous medium. The velocity of the magma through the crystalline matrix v_i , is given by Darcy's law (Bear, 1972)

$$Xv_i = -\frac{k}{\eta_i} \frac{dp}{dz} = \frac{k}{\eta_i} (\rho_s - \rho_l) g \quad (6.6.1)$$

where k is the permeability, η_i the viscosity of the magma, and dp/dz the vertical pressure gradient on the fluid. The vertical pressure gradient is assumed to be equal to the differential buoyancy of the magma and crystalline matrix, i.e., to the product of the density difference (ρ_l magma density, ρ_s crystalline matrix density) and the acceleration of gravity g . Implicit in this assumption is the negligence of any pressure difference between the fluid and matrix. At the high temperatures associated with partial melting, and on the relevant time scales, the crystalline matrix does not have sufficient strength to maintain a significant pressure difference.

In order to relate the permeability, k , to the porosity, X , a reasonable model is a cubical matrix of circular tubes; the matrix has dimensions, b , and the tubes have diameters, δ . This is illustrated in Fig. 6.6.1b. The resulting porosity is given by

$$X = \frac{3}{4} \pi \frac{\delta^2}{b^2} \quad (6.6.2)$$

And for slow (laminar) flow in the channels the permeability is (Bear, 1972)

$$k = \frac{X\delta^2}{9b} \quad (6.6.3)$$

Elimination of δ from Eqs. (6.6.2–3) gives

$$k = \frac{X^2 b^2}{72\pi} \quad (6.6.4)$$

The velocity of magma migration is obtained by substituting Eq. (6.6.4) into Eq. (6.6.1) with the result

$$v_i = \frac{Xb^2(\rho_s - \rho_l)g}{72\pi\eta_i} \quad (6.6.5)$$

It is implicit in this model for magma migration that the crystalline matrix can deform to accommodate the loss of the liquid fraction as it percolates upward due to its differential buoyancy. Solid-state deformation processes, either ionic or dislocation, are sufficient to provide the deformation.

It is of interest to evaluate the magma migration velocity for typical asthenospheric conditions. The matrix dimension is taken to be $b = 0.2$ cm. This is a typical grain size for mantle rocks. The density difference is taken to be $\rho_s - \rho_l = 0.6$ gm/cm³ and $g = 10^3$ cm/sec². The migration velocity is given in Fig. 6.6.2 as a function of the volume fraction of liquid for magma viscosities of 10, 100 and 1000 poises. A viscosity of 10 poises should be a reasonable value for a basaltic magma in the asthenosphere and andesitic magmas should have viscosities near 10^3 poises (Kushiro *et al.*, 1976; see also section 9.4.3). Typical mantle flow velocities are of the order of 1–10 cm/year. It is seen from Fig. 6.6.2 that the magma migration velocities are much larger than this when the liquid fraction is between 1–10%. Our conclusion is that magma will percolate upwards through the asthenosphere as rapidly as it is produced, although some recent studies (e.g., Waff, 1980; Walker *et al.*, 1979) suggest that relatively small magma bodies may not escape. The liquid fraction present will be close to that required to produce interconnected channels (that is, a few percent). At small percentages of partial melt the rate of magma flow is strongly dependent on the fraction of melt present, according to Sleep (1974).

An important question to address is whether the magma, as it percolates through the asthenosphere, will coalesce to form larger bodies of magma. In order to move through the crystalline matrix, these larger bodies would have to deform the matrix. It is appropriate to consider the motion of a spherical body of magma of diameter d . The crystalline matrix may be modelled as a Newtonian fluid with a solid-state viscosity η_s . Since this solid-state viscosity is much larger than the magma viscosity η_l the velocity at which the magma body will rise, v_i , is given by Lamb, 1945 as

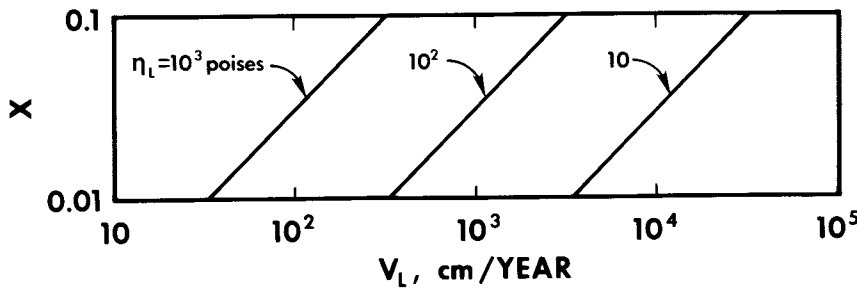
$$v_l = \frac{d^2 g (\rho_s - \rho_l)}{12 \eta_s} \quad (6.6.6)$$

This velocity is given as a function of the diameter of the body in Fig. 6.6.3 for several values of the solid-state viscosity and other parameters as previously given. From studies of postglacial rebound, a reasonable value for the solid-state viscosity of the asthenosphere is $\eta_s = 4 \times 10^{20}$ poises. Because of the resistance of the crystalline matrix to deformation, the buoyancy-driven velocity of small magma bodies is infinitesimal. The percolation along the intersections of grain boundaries is a much more efficient mechanism for magma migration and it is likely to dominate in the asthenosphere.

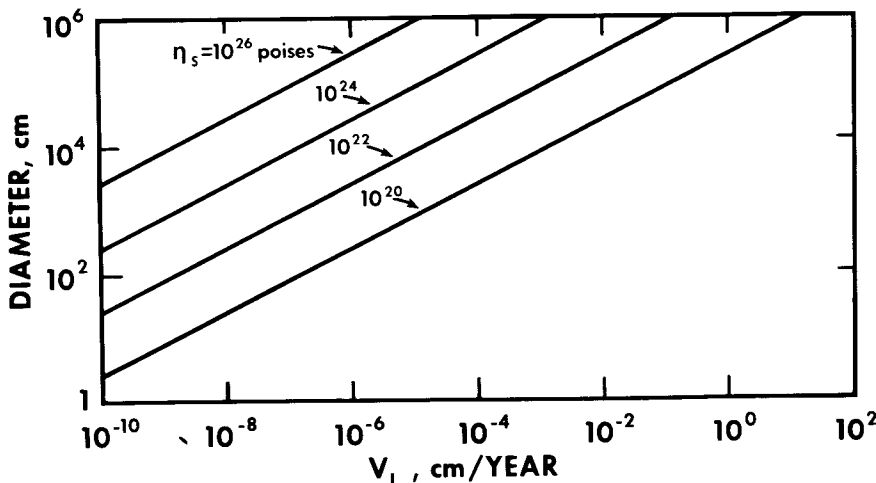
Under the influence of gravity, magma is expected to rise vertically through the asthenosphere until it reaches the lithosphere. At oceanic ridges the lithosphere is thin and this flow is likely to reach the magma chamber, which is believed to lie beneath the central ridge crest. In other areas of volcanic activity, such as island arcs or intraplate volcanism, the lithosphere is considerably thicker and the magma is expected to collect at much greater depths. The magma that has migrated upward through the asthenosphere would be expected to form a liquid layer at the base of the

lithosphere. A number of authors (i.e., Marsh and Carmichael, 1974) have proposed that the collected liquid will penetrate the lithosphere diapirically. The collected magma will certainly be lighter than the overlying lithosphere and a Rayleigh-Taylor instability will result if the lithosphere exhibits a fluid-like behavior. Laboratory studies of the diapiric penetration of a light, low-viscosity fluid have been carried out by Whitehead and Luther (1975). They found that nearly spherical bodies of the lighter fluid rose through the heavier fluid. This mechanism of magma migration is illustrated in Fig. 6.6.4a.

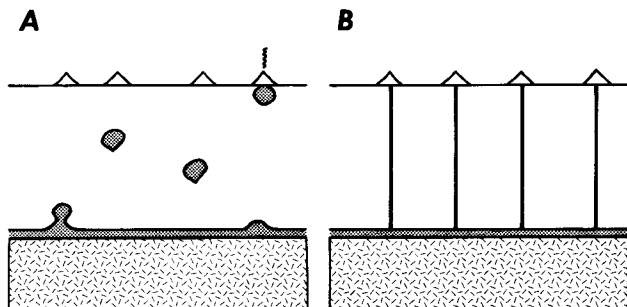
A model for the diapiric penetration of magma through the lithosphere is the vertical rise of a spherical magma body through a Newtonian viscous medium. The velocity at which such a body will rise has been given by Eq. (6.6.6). For lithospheric deformations a viscosity of 10^{24} – 10^{26} poises is usually assumed. For these viscosities the velocities are extremely small as can be seen in Fig. 6.6.3. The transfer of heat from the magma body may significantly reduce the equivalent viscosity of the rock through which the rock is moving. The coupled problems of flow and heat transfer have been considered by Marsh (1976, 1978).



6.6.2 Migration velocity of magma (V_L) as a function of volume fraction of liquid (X).



6.6.3 Velocity at which a magma body will rise (V_L) related to its diameter (d) for four values of the solid-state viscosity of the mantle.



6.6.4 Magma migration. In **a** nearly spherical bodies of a lighter fluid rise through a denser fluid (as Whitehead and Luther, 1975, found experimentally) and in **b** the magma rises in continuous conduits.

An alternative model for the migration of magma through the lithosphere assumes that continuous conduits develop. These conduits act as pipes through which the magma flows. The model is illustrated in Fig. 6.6.4b. A volcanic conduit can be modelled as a circular vertical tube of diameter d . If it is assumed that the differential buoyancy is driving the flow, the vertical pressure gradient on the magma is $(\rho_s - \rho_l)g$. If the flow in the tube is laminar, the mean magma velocity is given by Schlichting (1955)

$$\bar{u}_l = \frac{d^2(\rho_s - \rho_l)g}{32\eta_l} \quad (6.6.7)$$

If the diameter of the tube is sufficiently large the magma flow becomes turbulent. The transition occurs when $\rho_l \bar{u}_l d / \eta_l \approx 3000$. Substitution of Eq. (6.6.7) gives the critical tube diameter

$$d = \frac{45.78\eta_l^{\frac{3}{2}}}{[\rho_l(\rho_s - \rho_l)g]^{\frac{1}{2}}} \quad (6.6.8)$$

If the diameter is greater than this, then the flow is turbulent and an empirical expression for the mean velocity is (Schlichting, 1955)

$$\bar{u}_l = 0.349 \frac{(\rho_s - \rho_l)^{\frac{4}{3}} g^{\frac{2}{3}} d^{\frac{5}{3}}}{\rho_l^{\frac{2}{3}} \eta_l^{\frac{1}{3}}} \quad (6.6.9)$$

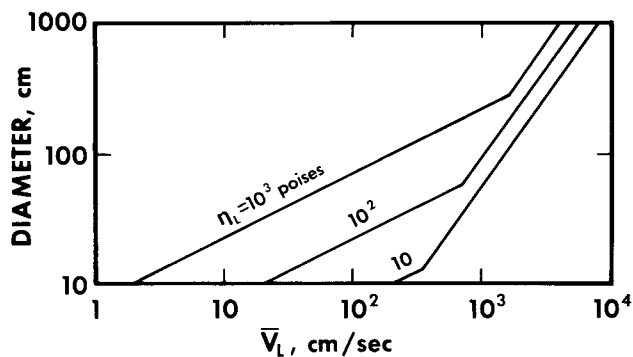
For turbulent flow the mean velocity is a very weak function of the magma viscosity. In Fig. 6.6.5 the mean magma velocity is given as a function of the tube diameter for three values of the magma viscosity ($\rho_l = 2.7 \text{ gm/cm}^3$, $\rho_s - \rho_l = 0.6 \text{ gm/cm}^3$, $g = 10^3 \text{ cm/sec}^2$). The mean velocities are in general quite large.

Fujii and Uyeda (1974) have suggested that the coupling between thermal dissipation and a strongly temperature-dependent viscosity will lead to flow insta-

stabilities in volcanic conduits. In a system with a constant driving force it has been shown (Gruntfest, 1963) that the decrease in viscosity caused by viscous heating can lead to flow instabilities. Hardee and Larson (1977) have questioned whether this mechanism can be responsible for the episodicity of volcanism. They doubt that the temperature dependence of the viscosity in a magma is sufficiently large over a sufficient temperature range to result in this type of instability.

It has been suggested that a fraction of the buoyancy head is required to build volcanoes to their observed height (Vogt, 1974; Edman, 1976). If this is the case, then conduits through the lithosphere are required.

Near-surface observations indicate that magma is capable of fracturing brittle rocks. A number of authors have considered this problem (Roberts, 1970; Pollard, 1973; Pollard and Muller, 1976). The wall cooling of a magma flowing through a dike has been considered by Maaloe (1973). It has been suggested by Weertman (1971) that magma may propagate through the lithosphere by hydrofracturing. Anderson and Grew (1977) have suggested a corrosion crack propagation mechanism.



6.6.5 Mean magma velocity (\bar{V}_l) as a function of diameter of a conduit as illustrated in Fig. 6.6.4b for three values of magma viscosity.

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