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Peter Bird
Department of Earth and Space Sciences
University of California
Los Angeles, California 90024

8

TECTONICS OF THE TERRESTRIAL PLANETS

ABSTRACT

Because geologic events are not reproducible, the unraveling of past tectonic events is always difficult and controversial. This is especially so for other planets, where the available data provide even less complete knowledge than we had of the Earth in 1930. Because of these limitations, we cannot study planetary tectonics empirically, so we are limited to testing for the presence of particular styles suggested by theory and intuition. The major possibilities are nonuniform contraction, despinning, volcanism, isostatic adjustment, and several varieties of convection, including homogeneous, layered, plate-tectonic, delaminating, and plumose.

All of the five planets considered show volcanism and some isostatic adjustment in their early histories, supporting the concepts of hot accretion, early differentiation, and gradual cooling. On the smaller planets (the Moon and Mercury), a global lithosphere formed early and remained unbroken or was only slightly cracked by the small strains of nonuniform contraction and perhaps despinning. On Mars, the plains and highland provinces may record a phase of homogeneous convection, with drift of blocks of primary crust. Now there is a global lithosphere, disrupted by the Tharsis plume, which thinned it from below, and fed volcanism, which loaded the thick lithosphere until it was bent and cracked on a global scale. The Earth also has had a history of waning convection and growing plates, with two important changes in style. After the first eon, continents became too big and buoyant to be recycled through zones of downward flow. After the Archean Era, the convection probably changed from the symmetrical homogeneous to the asymmetrical plate-tectonic form. The retention of a hydrosphere here is responsible for most of the present volcanism, the formation of continental crust, the thickness of oceanic crust, their neat separation at two different elevation levels, and the asymmetry of subduction. Our expectation that higher surface temperatures on Venus should cause even greater tectonic activity is confounded by radar images showing circular features like ancient impact craters. This observation seems to require a Martian model with a global lithosphere, where the plateaus known as "*terrae*" are underlain by plumes. Such a model can be reconciled with rock mechanics if Venus has no significant granitic crust. However, if future data show large areas free from craters, and a granitic composition in the terrae, then Venus would more probably be a unique example of layered convection, with complementary overturning cells in both the crust and mantle. The lithosphere might then be thinner than the crust, and consist of a set of thin, brittle "rafts" which preserve some ancient surface without impeding the convection below.

8-1 INTRODUCTION

Tectonics is the science (or art) of using observations made across a two-dimensional surface during one "instant" of geologic time to infer a complete history of the three-dimensional strains and movements of the rocks. Clearly, the data are inadequate even when rocks are perfectly exposed and observed, so many constraining assumptions are needed. A popular guiding principle is that of greatest simplicity (Occam's razor), which holds that structures are assembled from a few homogeneous rock types and become more complex through time. However, believers in uniformitarianism are forced to accept complementary simplifying processes (e.g., erosion, melting). Once these are allowed into a reconstructed history, any appearance of a unique solution is destroyed. Clearly, a purely empirical approach to tectonics is not enough.

The empirical method is even weaker when the data are limited to what can be observed from orbit and a handful of landing sites (see Table 8-1). Collecting images with available light or active radar reveals the broad contours of the surface and such details as fault lines, lava flows, and craters. But little or nothing can be inferred about rock composition, and the fine textures by which a field geologist would infer the youth or age of a landform are not resolved. In fact, the technique of crater counting (see boxed material on p. 87) is the only way of dating rocks we have not sampled, and it does not work well for recent epochs when the meteorite influx was slow. Gravity measurements yield smoothed estimates of the minimum-possible stress and anomalous-mass distributions, but there is no assurance that the planet adheres to a minimum principle, or that the long-wavelength parts of its tectonics are the most important. A seismicity sample may be obtained, but over a week to a year the observed pattern is apt to be dominated by one or more aftershock swarms, rather than the true strain-rate distribution in the planet.

TABLE 8-1 Available Data on Planetary Tectonics

Method	*Reveals*	*Limitations*
Imaging	Crater-count ages, faults, volcanic deposits	About 1 km resolution, only relative elevations
Radar altimetry	Major structures and loads	About 30-km resolution
Satellite gravity	Degree of isostasy	Wavelengths $\geqslant$ orbit height
Seismicity	Lower limit on fault slip rates	Short time sample, poor or no event locations
Heat flow	Lithosphere thickness	Questionable accuracy, limited to Lander sites
X-ray fluorescence	Major-element bulk chemistry of crust	Limited to Lander sites

To make such limitations more graphic, consider what we would know of the Earth if we had investigated it from another planet with *Mariner* and *Viking* technology. We would probably extrapolate coastal topography beneath the oceans, never guessing their true depth or the existence of a second (basaltic) type of crust. We would detect subduction-zone volcanoes, a few (mostly inactive) faults, and only one impact crater. The most visible tectonic belts would be the Alpine-Himalayan continental-collision belt and the Rocky-Andes Mts. belt of the Americas, whose formation is somehow related to plate tectonics, but still not fully understood by our geologists on the ground. The landers sent down to smooth and featureless "safe" sites like the Amazon basin and Sahara might report such odd crustal compositions as hematite-bauxite or pure quartz, respectively. Neither lander would be likely to detect earthquakes with a short-period instrument on a windy site, and a Saharan lander might not even detect life! Even allowing for gravity and magnetic measurements from orbit, our knowledge of Earth would be far less than terrestrial geologists had available in 1930. Recall that it was another 35 y after that before the main (plate-tectonic) elements of Earth's orogenic cycle were discovered by empirical means.

Therefore, extraterrestrial tectonic models must still be motivated by theories more than by observations. By combining physical laws and laboratory observations on rock deformation, we can predict alternative classes of possible behavior of planetary surfaces. The suite of models can never be narrowed to a single prediction, because any theory will contain as parameters the temperature and chemical distributions in the interior, two things that are inherently unmeasurable even from the surface. However, a few scanty observations of actual faulting, gravity, and crater counting may then be put to powerful use as "filters" that reject whole classes of models and further focus our thoughts and future exploration.

The approach in this chapter is to summarize very briefly the relevant facts of rock mechanics and the classes of tectonic models to be found in the literature. For each, we emphasize specific predictions of observable features. Next we review the evidence on the terrestrial planets one by one and relate it to these models. In the end, we see what appears to be an excellent inverse correlation of tectonic activity with lithosphere thickness, which is in turn controlled by radius, age, and atmosphere.

However, the apparent orderliness of this approach does not preclude conclusions that are wrong. The theoretical approach may outrun the empirical, but it can also go astray if not all of the relevant physical processes have been considered. This has happened twice in the history of geology. First, William Thompson Lord Kelvin "proved" with a conductive heat-flow model that the Earth was only 20 to 40 My old; neither Lord Kelvin nor any other scientist of the nineteenth century knew of radioactive heat production, and few thought of convection as relevant. In this century, seismology "proved" that continental drift was impossible because the Earth's entire mantle was solid, and solid rocks could not flow; it was only a few decades ago that electron microscopes began to show us dislocations in crystals which prove that rocks do flow. In an audacious review such as this, there is almost certainly at least one fundamental error of that type!

8-2 ELEMENTS OF ROCK MECHANICS

To understand tectonics, it is essential to understand at least qualitatively the ways in which rocks can deform. This brief review is focused on physical mechanisms, rather than descriptive terms for the resulting structures. The most important mechanisms are (1) thermal expansion, (2) elasticity, (3) frictional faulting, and (4) dislocation creep. Each of these is a different type of *strain* (a word that is used here qualitatively to indicate the fractional change of shape of a body of rock). All four mechanisms may act at once under the same conditions, superimposing their strain effects. A related measure is the speed at which they operate to accumulate strain; we refer to this as the *strain rate* and measure it in units of (dimensionless) strain per unit time.

The causative factors leading to strain are primarily temperature and stress; some simple definitions regarding the latter are useful: *Stress* is the measure of how strongly equal and opposite opposed forces act on the opposite sides of internal planes in the rock. In the Systeme International (SI) metric system its unit is the Pascal (Pa), or associated unit megaPascal (MPa $= 10^6$ Pa). Because these forces have a direction as well as an intensity, we distinguish further between *normal* stresses (those with force acting normal to the plane in question) and *shear* stresses (those with forces acting parallel to the plane in question). It is useful to remember that fluids will relax (by straining) until they have no shear stress within them and until the normal stress in all directions is equal (we then call it *pressure*). However, solids may be much stronger, so that the shear stresses (or the differences in normal stress across different planes) must reach high levels to cause visible strain.

8-2-1 Thermal Expansion

When subjected to an increase of temperature, rocks undergo a fractional expansion of all their dimensions of 5 ppm to 12 ppm per degree Celsius. If the rock is more or less rigidly constrained, this may change the stress state and cause other types of strain. When rocks are weakly confined, they expand, and the important effect is the decrease in their density, by 15 ppm to 36 ppm per degree Celsius. Rocks that are hotter than their surroundings tend to rise because a reduced density implies *relative* buoyancy; this effect is what causes convection. The high pressures of planetary interiors reduce these thermal expansions by 70% or more but cannot eliminate them. The fundamental thermal-expansion effect underlies all but one of the tectonic models proposed below.

8-2-2 Elasticity

Elastic strain is, by definition, proportional to stress; therefore it returns to zero when stress is removed. Elasticity of rocks is also universal, but not important in tectonics. This is because the shear stresses that rocks can support (less than 10^9 Pa) give rise to strains of less than 1 percent, which are not visible or detectable by any type of remote sensing. Of course, the isotropic elastic compression of rocks by interior pressures affects planetary radii and densities, but these constant and spherically symmetric effects cause no surface deformations.

8-2-3 Faulting

The dominant deformation mechanism at the surface of any of the terrestrial planets is *faulting*, the sliding of one rock mass over another along the planar or curved surfaces known as *faults*. Faulting is independent of time and temperature, and its governing law states simply that the shear stresses (σ_s) acting on the fault must reach the critical value of $\sigma_s = \mu(\sigma_n - P)$ for fault slip to continue, where σ_n is the stress normal to the fault plane, P is the pressure of interstitial fluids (if any), and μ is the dimensionless coefficient of friction, approximately 0.8 ± 0.1. Remarkably, this law is valid regardless of rock composition (excepting hydrated clays), texture, strain rate, or temperature (up to 400°C), or pressure (up to 10^9 Pa).

In this chapter, we describe the geometry of faulting loosely as belonging to one of three categories observed on Earth. The angle between the fault plane and the horizontal (called the *dip*) takes one of three values: 30°, 60°, or 90°. *Thrust* faulting is the overlapping of one block onto another by slip on a fault surface that dips gently (30°) into the interior; it is caused by horizontal compression in the direction of sliding. *Normal* faulting is the uncovering of one block by another, with slip on a fault surface dipping more steeply (60°); it is caused by horizontal tension in the direction of sliding. Finally, *strike-slip* faulting is purely horizontal movement across a vertical fault plane; it is caused by horizontal compression in one direction combined with horizontal tension at right angles to it. Because an old fault full of crushed rock has slightly lower strength than other adjacent planes, old fractures tend to keep sliding even if stress directions change slightly; this leads to large cumulative offsets with linear topographic expressions. On Earth, the presence of water causes faults to anneal and stick together with time, so that when they finally move they must "refracture" with a jerk, and seismic waves are radiated (Angevine et al., 1982). On other planets where water is absent, it is not clear that fault slip necessarily produces detectable seismic events!

8-2-4 Dislocation Creep

Ironically, the faulting phenomenon gives us visible evidence of tectonic activity, but no tectonics would be possible if the friction equation cited above were universally obeyed. This is because the increase of normal stress with depth would make the necessary shear stresses impossibly large. Therefore, we infer that in every planet there is an additional (weaker) strain mechanism: intracrystalline dislocation creep (Weertman and Weertman, 1975). This process strains rocks aseismically and without changing their volume. Strain continues to accumulate as long as shear stress acts, at a rate given by:

$$\dot{e}_s = A\sigma_s^3 \exp\left(\frac{-E - PV}{RT}\right)$$

where $\dot{e}_s$ is shear strain rate, R is the gas constant, T is absolute temperature, P is pressure, and the three constants A, E, and V depend upon the mineralogy. On Earth the important creeping minerals are quartz (in continental crust), which softens significantly above 400°C, and olivine and pyroxene (in the mantle), which creep significantly above 800°C. The activation energies (E) in each case are so high that strain rate (at constant stress) more than doubles for each 20°C rise in temperature. Conversely, the strain rate is at least halved for every 20°C drop, and becomes undetectable below the temperatures stated.

8-3 TECTONIC MODELS

Kaula in Chapter 4 has summarized the formation and structure of the terrestrial planets, and has concluded that each of the inner five has a thick mantle layer in which olivine or pyroxene is important. Therefore, we shall boldly assume that within each planet the same deformation mechanisms operate and that there is a transition at about 600 ± 200°C between the mechanism of faulting and that of creep. The layer above this transition is called the *lithosphere*, which consists of a single spherical shell, or a set of spherical caps (separated by faults); heat transfer within it is by conduction, and the vertical temperature gradient is large. Below the transition is the *asthenosphere*. With depth, this creeping layer rapidly loses strength until convection begins; the temperature gradient is thereafter small (nearly adiabatic) except in boundary layers that form around any barrier to vertical flow. The depth of the lithosphere-asthenosphere boundary is a function of the strain rate, sinking when the rocks are deformed quickly, and rising in quiescent periods. But in foreseeable cases, it would never rise to the surface or sink to the core. This basic division forms the conceptual basis for the following nine possible mechanisms that might deform or modify a planetary surface.

8-3-1 Nonuniform Contraction

Some overall temperature change is inevitable in the history of a planet, and except in the case of a special coincidence, thermal expansion (or contraction) of the interior does not match that of the surface. Within the interior, creep will relax any shear stresses quickly, leaving only a uniform anomaly in the pressure (with respect to lithostatic pressure). In turn, this interior pressure anomaly causes deviatoric stress in the thin, stiff lithosphere. These horizontal tensions (or compressions) in the lithosphere are greater in size than the pressure change, as the stress in a balloon exceeds the gas pressure within it. Stresses are amplified by approximately the ratio of planetary radius to lithosphere thickness. If that ratio is 100 (as in the Earth), any differential temperature change of 25°C between lithosphere and asthenosphere results in stresses exceeding the frictional-sliding limit of the lithosphere. The likely result would be a distributed set of faults through the lithosphere, in a variety of directions, with a spacing determined by the ability of the lithospheric blocks to withstand secondary internal stresses caused by the change in planetary radius (see Fig. 8-1a). If the contraction of the interior were greater than that of the lithosphere, these faults would be thrusts, dipping about 30° and pushing up ridges. The opposite situation would form normal faults, dipping 60° and dropping down linear blocks to form valleys. The height of these topographic features would depend upon the small difference in strength between faulted and unfaulted lithosphere and is thus very difficult to predict.

8-3-2 Despinning

The lithosphere of a planet may also cease to "fit" if the planet's rate of rotation changes, because this changes the degree of ellipticity required for a *lithostatic* (shear-stress-free) state. If inelastic tidal strains slow the rotation, the radius of curvature of the lithosphere will be forced to decrease near the poles and increase near the equator.

Melosh (1977) has analyzed this complex problem and found that when a planet has a thin lithosphere, the lithosphere strains elastically and does very little to retard the change in ellipticity. The resulting stresses are then dependent only on the ellipticity changes and not on lithosphere thickness or depth. This means that the upper parts fault, whether or not these faults cut through the entire lithosphere. The resulting faults are east-west-trending normal faults at high latitudes, intersecting strike-slip faults trending N60E and N60W in the midlatitudes, and (possibly) north-south-trending thrust faults near the equator (see Fig. 8-1b). Naturally, an admixture of nonuniform contraction can

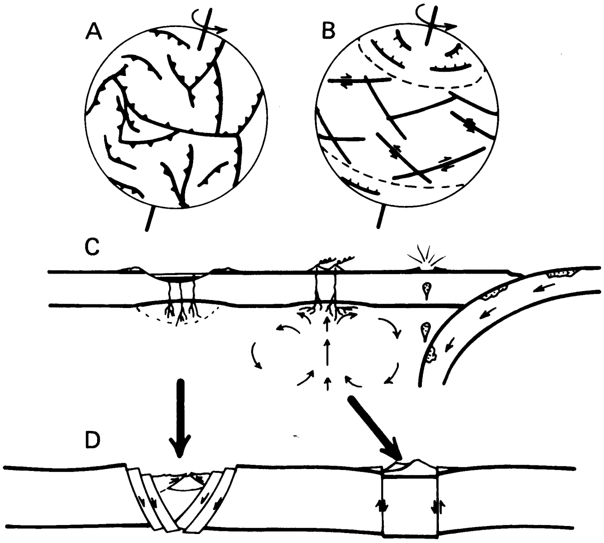

Fig. 8-1 Four types of planetary tectonics that are not primarily convective: (A) Nonuniform contraction produces a global fault network with no preferred orientation as the lithosphere is expanded or contracted to fit; (B) despinning produces conjugate strike-slip faults in the equatorial zone and normal faults near the poles; (C) volcanism can result from pressure-release melting after impact cratering (left) or during convection (center) or because crustal material or volatiles are convected back to the interior (right); (D) isostatic adjustment through faulting may involve slumping to fill a crater (left) or cylindrical faulting to adjust to volcanic loads (right).

change the pattern by expanding either the thrust-faulting or the normal-faulting regions. However, the predominance of east-west compression over north-south compression at all latitudes remains, and the resulting global orientation pattern of fault trends indicates that despinning has occurred.

It should be obvious that both of the mechanisms above operate very slowly, so their effects will not be noticed if the planet surface is modified by the faster processes below.

8-3-3 Volcanism

Melting of the interior is not something that a theoretician would predict, given a planet-sized lump of uniform, monomineralic rock. Instead, one would expect that slow heating from radioactivity and other causes (see Chapter 12 by Peale) would raise internal temperatures until the strong thermal activation of creep enabled convection to remove heat just as fast as it was produced. Similarly, it has been shown that steady fault slip cannot melt rocks, because of the thermal-softening effect (Yuen et al., 1978). Therefore, melting requires special circumstances, of which three classes are presently known.

The first is melting due to large meteorite impacts, which were probably far more common before 4 Gy. Contrary to intuition, the impact itself produces limited melting of the crater floor, because the total melt produced in escape-velocity impacts with terrestrial planets is only about twice the volume of the meteor (O'Keefe and Ahrens, 1977). Furthermore, most of this melt is splashed out along with the solid ejecta. The more important effect is that crater excavation immediately reduces the pressure on the mantle beneath the floor. Since the melting points of rocks are elevated by pressure, this sudden drop could melt asthenosphere that was already hot. According to this analysis, crater flooding may be taken as a rough indication that the lithosphere thickness at the time of impact was less than one crater diameter (see left-hand side of Fig. 8-1c).

The second melting mechanism relies upon volatiles (e.g., H_2O, CO_2) that are more soluble in melts than in solid rocks. If these are present (see Chapter 7 by Fanale), the melting point is depressed (without much effect on the creep strength), and the convective cycle may involve limited melting in the hot upwelling zones (see center of Fig. 8-1c). This is believed to be the cause of midocean-rise volcanism on Earth. Generally, the presence of such volatiles implies either incompleteness of planetary differentiation or convective overturning that returns volatiles to the interior from the atmosphere or ocean. The Earth presents an example of volatile recycling.

Third, a differentiated planet with a crust that was formed from a low-melting fraction of the mantle may be volcanic if convection draws small pieces of crust back into the interior. Although the geochemical evidence is mixed, it is possible that this effect contributes to subduction-zone volcanism on Earth (see right-hand side of Fig. 8-1c).

8-3-4 Isostatic Adjustment

The principle of *isostasy* is useful to use in approaching the study of vertical tectonic movements. It holds that any planet with a weak lithosphere (or none) tends to adjust itself until all its provinces "float" on the soft asthenosphere at their natural equilibrium level. These levels are lower where the crust is dense or thin, and higher where it is less dense or thicker. Technically, balance is achieved when all vertical columns everywhere have the same weight per unit area of base, as (hypothetically) measured at some common depth in the asthenosphere. It follows that every high mountain or plateau must be balanced by low-density (hot) mantle or an extra thickness of relatively low-density crust below it. These possibilities are illustrated in Fig. 8-4 in Sec. 8-4-3 later in this chapter. These low-density bodies at depth are referred to as the *compensation* of the surface topography, which is said to be *perfectly compensated* when isostasy holds exactly.

However, isostasy resembles a human law more than a natural law, in the sense that it is an ideal from which there are many departures. On Earth, for example, we often find that features either less than 30 km in width or less than 10^7 y old are partially to totally uncompensated. The explanation is that a strong lithosphere can violate isostasy and hold topography above or below its isostatic level. An important part of tectonics is the study of the deformation of the lithosphere by vertical loads arising from lack of perfect compensation. These loads are often thought of as positive (as when volcanism adds mass to a planet's surface), but they may also be negative (as when cratering or erosion remove mass). Finally, a net load may be created if mantle convection changes the subsurface compensation without changing the topography.

In the absence of subsurface seismic-structure information, it is difficult and controversial to determine the true size of these loads. An integration of the topography fails to account for the anomalous subsurface densities that often reduce the net load, so it would seem that gravity would give a better indication. On the other hand, the load indicated by the integral of the gravity anomaly depends upon the part of the surface included in the integration and vanishes when this surface includes the whole planet. Hence any interpretation must be based on a geologic model and remains subject to revision. Specialists in this field attempt to distinguish between alternative models by studying how and why the ratio of gravity anomaly to topography varies with wave-length.

As soon as the load is imposed, dislocation creep at all levels begins to relax shear stress locally, so that it must be carried by shallower, colder levels of the lithosphere. This process decelerates but never ends, so theoretically all cases that are not disrupted go through an initial "elastic-bending" phase followed by faulting, collapse of the topography, and unbending. On Earth we have ancient shorelines establishing convenient datum surfaces with which to study the bending and infer lithospheric thicknesses. On other planets, only the second phase that produces visible faults is open to study.

These faults can be classed into two groups according to the direction of rock movement. The first consists of cylindrical faults with a vertical slip direction that surround the load and drop (or raise) the inner topography closer to its isostatic level. The second set occur only if the central region is hot and weak, a mechanical "hole" in the lithosphere. In this case, the anomalous pressure existing between the shallow load and its deeper compensation also acts horizontally to drive the lithosphere radially away from topographic highs (into lows). The associated map pattern of faults may include circumferential thrust (normal) faults, radial (normal) thrust faults, conjugate logarithmic spirals of strike-slip faults, or a combination, depending on local topography and temperature patterns (Fleitout and Froidevaux, 1982). Such fault patterns may be expected to spread radially outward at a decreasing rate until the central topographic or thermal anomaly disappears. However, if the lithosphere is thick, this process may take much longer than the present age of the solar system.

One final point about isostasy is that where there is volcanism, it may help a distant observer to estimate the thickness of the lithosphere or at least to put a lower limit on it. The argument is that the hot magma within a volcano is weak and will behave according to fluid laws. In particular, it will rise no higher than its isostatic level. If we take the top of the highest volcano to be the isostatic level of the magma, and the height of the surrounding plains to be the isostatic level of the cold lithosphere, then the difference between them (Δh) is just given by

$$\Delta h = T\left(\frac{\rho_L - \rho_M}{\rho_M}\right)$$

where T is lithosphere thickness, ρ_L is lithosphere density, and ρ_M is magma density. If we make a further assumption that the magma and lithosphere are of similar chemical composition, we can apply terrestrial values of 4 to 10 percent for $(\rho_L - \rho_M)/\rho_M$ (Daly et al., 1965) to other planets. Then T will be from 10 to 25 times the size of Δh.

8-3-5 Convection

Current theoretical models of the thermal histories of the terrestrial planets (Toksöz and Johnston, 1977) show that each should now have an asthenosphere in active convection, regardless of the planet's formation temperature. However, this review concerns only those convection systems that break and displace the surface, allowing remote sensing. Classical convection theory is of limited help, as it is just beginning to address the three major complications that may restrict planetary convection to the interior. One is the differentiation of a crustal layer whose compositional buoyancy overwhelms the thermal density anomalies tending to cause overturn. The other complications are the extreme temperature dependence of rock strength and the extreme nonlinearity of the creep law (i.e., lower effective viscosity at high stress); both of these make planetary convection more episodic and bistable than classical linear models. That is, substantial potential energy may accumulate before some large perturbation allows surface convection to begin; thereafter convection may be self-sustaining for some time by providing the heat and stress that keep surface viscosities low; and when it has dissipated the potential energy, it may "freeze up" and stop (or change form) with equal suddenness. Therefore, in searching for evidence of the following five forms of surface convection (see Table 8-2), we should not expect to see them in action throughout the whole surface or the whole history of a planet.

TABLE 8-2 Types of Planetary Convection

	Requires	*Produces*
Homogeneous	Hot formation or radioactivity	Young or mixed-age surface, lineated or polygonal topography
Layered	Global crust of low viscosity	Young or mixed-age surface, lineated or polygonal topography
Plate-tectonic	Strong boundary layer, fault-weakening	Young or mixed-aged surface, lineated topography with assymetrical trench-arc convergence zones, strike-slip faults
Delaminating	Crust with low-viscosity basal layer	Migrating regional uplifts (surface may be old)
Plumose	Heating from below (core, layered mantle)	Local uplift of spot or ramp (surface may be old)

Homogeneous convection *Homogeneous convection* is the type observed in laboratory experiments on homogeneous fluids. It could only occur in a planet if the volume of crust were small and if the formation of a strong upper-boundary layer were suppressed (by a high surface temperature or some nonthermal weakening mechanism). Upwelling and downwelling are both locally symmetrical, so the flowlines may form closed loops in a pattern with long-term stability. Experiments show that a pattern of horizontal rolls converts to a pattern of hexagonal cells as convection gets more intense; however it can safely be predicted that even a weak surface lithosphere (which resists stretching more than bending) forces rolls to be dominant in the case of the planets. Thus, a lineated topography with alternating ridges and trenches in a young surface is produced (see Fig. 8-2A). The gravity field in this case is difficult to predict, because the primary convective mass anomalies cause deflections of the surface with an opposing gravity effect; these deflections have less excess mass per unit area but are closer to any observer. All that can be predicted is that the gravity anomaly is less than (or opposite from) that produced by the topography alone.

If a small amount of surface crust were added to this system, it would be swept into the downwelling depressions, but its relative buoyancy would presumably keep it on the surface. This would result in small patches of old surface surrounded by young surface.

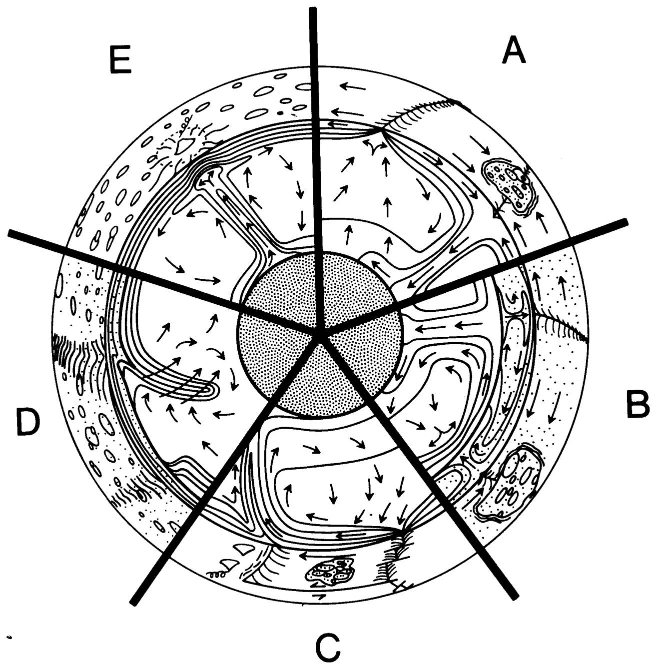

Fig. 8-2 Five types of planetary convection. Inner ring is the core, intermediate ring is a section through the mantle, and outer ring gives an oblique view of the surface. Solid curves are isotherms, and arrows indicate flow. Crustal material is dotted. Circle pattern on surface indicates impact craters and ancient surface. A = homogeneous, B = layered, C = plate-tectonic, D = delaminating, and E = plum convection. All cases except E are heated by radioactivity in the mantle; E is heated from the core below.

Layered convection *Layered convection* results when the amount of crust is increased to form a planetwide layer, but the lithosphere thickness is much less than the crustal thickness. Now the pattern of horizontal rolls appears in each layer, and the requirement of continuity of velocity across the boundary produces an out-of-phase pattern (see Fig. 8-2B). The upper, crustal layer is cooler, but does not necessarily have higher viscosities, since density, melting temperature, and creep-softening temperatures are correlated in rocks. Independent of viscosity, the thin upper layer would be less unstable than the thick mantle, and its convection would be "passive" (i.e., driven by shear stresses transmitted across the boundary); therefore the width of the cells would probably be determined by the mantle layer. At the surface, this case would look like homogeneous convection, except that the surface rocks would not be dense enough to account for the planet's known mass and inertia values without invoking an unusually large core. The prediction of gravity for this case is impossible without a number of constraints, because there are two layers of (opposed) thermal anomalies and two interfaces with (opposed) deflections to consider—almost *any* ratio of gravity to topography could result. As in the case of homogeneous convection, there might be "rafts" of lithosphere preserving patches of old surface above the downwelling zones.

Plate-tectonic convection *Plate-tectonic convection* is a style no one would have predicted. As we know it on Earth, it involves the formation of stiff, cold upper thermal boundary layers (lithosphere) onto which the crustal layers are strongly bonded by high viscosity. Two special circumstances make it possible for downwelling to occur anyway. First, the low-density crust of the continents is distributed unevenly about the globe (and is cold and stiff enough to stay that way), so that many downwelling zones can operate for a long time without encountering this "indigestible" substance. When they do, they stop, and new sites are formed elsewhere. Second, the basalt crust of the ocean basins undergoes a phase change to become a garnet-bearing rock called *eclogite* at 10^9 Pa pressure while it is being carried downward. This makes it even denser than the mantle below; hence it only hinders downwelling in the earliest stages and later promotes it (Ahrens and Schubert, 1975).

Plate tectonics shares the aspect of linear topographic features with homogeneous and layered convection. It differs in having long strike-slip faults along plate edges, which should be arcs of small circles. The general form of downwelling zones is less predictable, since such Earth-style features as volcanic arcs and fore-arc basins are not an essential part of the mechanism. There should, however, be a trench of some kind, with a consistent asymmetry caused by the fact that only one plate is being recycled into the interior (see Fig. 8-2C).

Significantly, no one has yet produced a laboratory model or computed simulation of convection in a homogeneous fluid that has the plate-tectonic style. In order to obtain great rigidity within lithospheric plates and also allow for relative movement of these plates, it is necessary to introduce heterogeneous bands of weak material (Schmeling and Jacoby, 1981). On Earth these bands are strike-slip faults and the thrust faults below the ocean trenches, which are many times weaker than plate interiors (Bird, 1978; Lachenbruch and Sass, 1980). The mechanical reason for this is not known, but most theories involve the physical or chemical effect of water in pores within the fault zones.

If such weak faults did not exist, the lithosphere on Earth would cease to move and would cool until it formed a single global shell capable of obscuring the convection below. However, there are still two ways in which such "buried" convection can be expressed at the surface—delamination and plume convection.

Delamination *Delamination* can occur if there are two lithospheres (on top of the crust and mantle layers, respectively) separated by a lower crust which is hot and viscous. Then, there may be a localized and rapid overturn beneath the crust that drops away the cold thermal boundary layer of mantle lithosphere and replaces it with hot asthenosphere from the mantle interior (see Fig. 8-2D). In order for this process to begin, some force must rift the lithosphere so that the asthenosphere can intrude, and then the lower-crust viscosity must be low enough to let the intrusion spread faster than it cools (Bird, 1979). Once it is underway, the surface expression is an elevated plateau with straight or gently curved boundaries, ringed by a deep linear depression (Bird and Baumgardner, 1981). The lateral growth of this plateau is at a rate of centimeters per year. There would be no deformation of the (old) surface, and subsequent cooling would return that surface to its original elevation in 10^7 to 10^8 y, so the detection of this process on other planets is more a theoretical than a practical possibility.

Plume convection *Plume convection* is another consequence of the temperature dependence of viscosity, this time at the hot and fluid end of the scale. If a planet has heat sources in its core, or chemical stratification in its mantle, the upper-mantle layer is partially heated from below (as opposed to being heated only internally by radioactivity). In this case, theory and experiment both predict that the hot lower thermal boundary layer is unstable, giving off hot "plumes" that rise rapidly and independent of any general circulation. Studies show that plumes widen slightly as they rise, thus entraining any of the surrounding mantle that they have heated and maintaining long-term stability (Yuen and Schubert, 1976). The lowered viscosity inside the plume allows rapid upflow (as in a pipe) with negligible cooling; such flows are so stable that the surrounding circulation of the whole mantle can tilt them by more than 45° before they break down and form a new path to the surface.

Plumes are important in planetary tectonics because they are a concentrated heat source of great power. Where they meet the lithosphere from below, it is thinned. This produces a surface uplift in addition to any caused by the low density of the plume itself. Melting and volcanism are possible, as the very hot plume extends itself to lower-pressure environments. A narrow volcanic edifice upon a broad swell is the likely result (see Fig. 8-2E). However, mantle circulation may cause the "hot-spot" to move in jumps, as mentioned above. In that case there could be a chain of isolated volcanoes sitting on a broad linear "ramp" formed by the combination of plume movement and thermal subsidence of the old inactive "hot-spots."

8-4 ANALYZING THE PLANETS

Having (in theory, at least) reduced the possibilities for planetary evolution to combinations of nine basic mechanisms, let us examine each planet in turn to see which styles have dominated. This travelogue is arranged in the order of apparent tectonic complexity, which is almost the same as the order of planetary radii (see Fig. 8-3).

8-4-1 The Moon

Of the terrestrial planets other than Earth, we have the most information about the Moon. Because some of the *Apollo* missions included landings, it was possible to measure surface properties (e.g., moonquakes and rate of heat flow from the interior) and to return samples for chemical analysis and dating. These are extremely valuable supplements to remote observation.

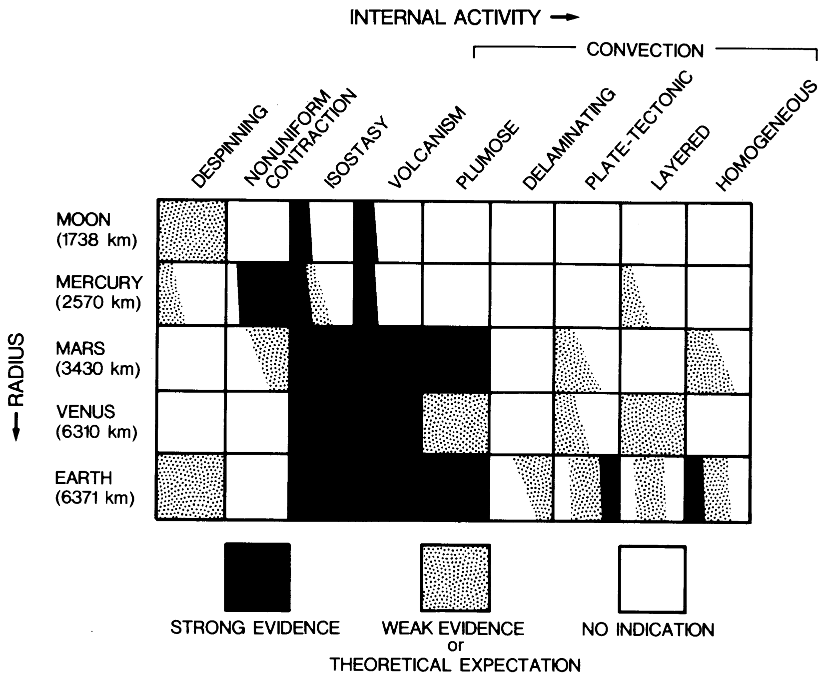

Fig. 8-3 Summary of inferred tectonic styles on the terrestrial planets. Where data permit, placement of shading and blackening in each box roughly indicates the portion of the planet's history concerned (left to right). Empty boxes do *not* indicate that a particular style is precluded by evidence, only that there is presently no indication that it has been operative.

The history of the Moon is actually less controversial than that of the Earth, because it involves relatively few provinces, they are perfectly exposed, and they are peppered with craters that give relative ages. The first known event is the segregation of a crust from an early "magma ocean" about 4.6 to 4.2 Gy (see Chapters 4 and 5 by Kaula and DePaolo, respectively). When this ocean began to freeze, the heavier minerals sank, but the lighter feldspar crystals (2.63 to 2.68 g $\cdot$ cm^{-3}) floated upward to concentrate in a global crustal layer about 60 to 100 km thick (Goins et al., 1977). Immediately the crust was subjected to heavy meteorite bombardment, not as a separate event but as a continuation of planetary accretion. It is likely that in the early stages this crust was hot enough at deep levels to creep and fill in the larger craters that cut deeply into it; this would explain the subdued form of the old craters on the side of the Moon away from Earth (e.g., the crater Mendeleev).

After the lithosphere had thickened to include the whole crust (about 3.9 Gy?) large impacts occurred, which made great holes in this crust to form the maria (see Chapter 5 by DePaolo). The first mystery about this event is why all the maria are on one side of the Moon. Did a single large planetesimal break up, in a pass close by the Earth, and then strike the Moon as a set of fragments? Did the Moon orbit close to the Earth with synchronous rotation, but with the present inner side formerly outward and exposed?

After these impacts, the craters were partly filled with low-viscosity basalt lavas, of density around 3.1 g · cm^{-3}. The principle of isostasy probably holds in the hot, disrupted region under any new crater, so the failure of this lava to completely fill the crater can be attributed to its density, which is significantly higher than that of the surrounding crust. The second problem is whether these basalts originated from pressure-relief melting under the craters, or whether a preexisting global layer of partial melt in the mantle (perhaps 200 km to 400 km down) was tapped to fill these holes. The first theory has difficulty explaining the 0.8-Gy duration of volcanism shown by crater patterns and radiometric ages (although it might take this long for an uplifted region of 200 km radius or more to cool down completely). The preexisting magma theory, on the other hand, has difficulty explaining just how the magma is moved and directed toward the crater at the proper time. Certainly, brittle tension cracks could not be opened at depths of more than a few tens of kilometers, due to the limited shear strength of all rocks.

Regardless of the magma source, the mare would initially have filled to an isostatic level. Later freezing and thermal contraction (volumetrically about 10 percent) would have lowered the mare surface, so that younger lavas (which continue to rise to the same absolute elevation) could continue flooding it with nonisostatic mass. Today we see *mascons* (concentrations of extra mass) over the maria, where the excess gravitational attraction of order 0.2 cm · sec^{-2} implies an excess basalt thickness of order 1.5 km, or a total basaltic fill of roughly 15 km. If the lithosphere included the whole crust as we have assumed, it would already have been thick enough at the time of the last eruptions (about 3.1 Gy ago) to support this excess mass.

This discussion of the lack of perfect lunar isostasy does not imply that there was no isostatic compensation or deformation following cratering. For one thing, the large-diameter basins were almost certainly more than 15 km deep at the first instant; since their gravity anomalies are not larger, there must have been some isostatic adjustment to the excess mass of late basalt flows. Second, Orientale Basin is surrounded by three great circular ring faults, which are downthrown on the inside, between outward-tilted blocks; this is the pattern predicted by Melosh and McKinnon (1978) in their model of a lithosphere slumping radially into a hole. Subtle mare ridges in Mare Imbrium suggest a similar structure buried by lavas (Guest et al., 1979). If mantle upflow into the larger craters was a common process, then associated pressure-release partial melting could account for the basalts, and it would not be necessary to postulate a buried layer of partially melted mantle.

Solomon and Head (1980) have analyzed this process of incomplete isostatic adjustment of mare, in terms of the bending stresses created in an elastic lithosphere loaded by mascons and overlying a soft asthenosphere. They show that the radial stress becomes most tensile at a certain distance from the maria, which coincides with the concentric lunar rilles. *Rilles* are long, sinuous valleys that may be created by normal faulting; as the crust is stretched, a narrow central block bounded by two parallel faults drops lower than its surroundings. The distance of the datable rilles from their associated mare provides a measure of lithospheric thickness. This thickness seems to have varied geographically from about 25 km to 75 km during the time span of 3.8 Gy to 3.4 Gy before present. In contrast, it had been essentially zero at 4.2 Gy ago.

If this rate of thickening had continued to the present, the lunar lithosphere would be about 150 km to 450 km thick today. However, we would expect the rate of thickening to decrease, for two reasons. First, the thickening lithosphere itself slows the rate of heat flow outward to space, because it depresses convection to greater depths in the interior. Second, the decreasing total heat flow from the planet may eventually approach the rate of heat production by internal radioactivity, so that a steady state with a constant lithosphere thickness is asymptotically approached. In fact, from the two measured heat-flow values (average 18 $mW \cdot m^{-2}$), we would estimate the Moon to have about a 100-km-thick global lithosphere today. The absence of distributed faulting of the nonuniform-contraction type indicates that lunar temperatures are now stable (Solomon and Chaiken, 1976). The lithosphere is strong enough to prevent internal convection from showing any surface effects. Since the Moon has lost any volatiles it had (see Chapter 7 by Fanale) and does not have a large core (or perhaps any core) to heat its mantle from below, there is no theoretical reason to expect melting and volcanism now or in the future. In terms of surface tectonics, it is a dead planet.

As a footnote, we should mention the exciting discovery of moonquakes at depths from 700- to 1000-km, whose location, timing, and occasional reversal of slip direction (!) all reveal them to be driven by monthly tidal stresses (Cheng and Toksöz, 1978). The nature of the mechanism that allows faults to slide at such high pressure is an important riddle in rock mechanics. However, these moonquakes do not indicate the extension of lithosphere (in the tectonic sense) to that depth, but only that the rocks in the asthenosphere have a Maxwell viscoelastic relaxation time of one month or more. That is, when stressed rapidly (by the monthly tides), these rocks do not have time to creep and behave as elastic or brittle. Yet under slow steady stresses they may well flow and convect. This is possible if their viscosities are above 10^{17} Pa-sec, a constraint easily satisfied if the asthenosphere of the Moon has the same temperature and viscosity as that of the Earth (10^{21} Pa-sec).

In fact, the existence of heat-flow measurements allows an interesting comparison to be made between the interiors of the Earth and the Moon. Laboratory and computer models of convection have led to the development of a number of proportionalities or "scaling laws" of great generality, and these may be applied to planetary convection as well. The ratio of interior viscosities of two planets can be shown to be roughly the inverse cube of the heat-flow ratio, multiplied by the ratio of their gravities. This analysis gives a lunar mantle viscosity about four times that of the Earth; which in turn suggests a temperature only 70 to 190°C lower in the mantle of the Moon. Thus, we believe that the Moon's asthenosphere is convecting beneath its rigid outer lithosphere, although this cannot be directly observed or verified. This convection must be remarkably close to steady state, because even a very careful analysis finds no evidence of nonuniform expansion or contraction since 3.9 Gy (Golombek and McGill, 1983).

8-4-2 Mercury

We know very little of Mercury except its average density, weak magnetism, and what was seen of the surface in the two near approaches by *Mariner 10.* Through careful analysis of the overlapping landforms in these images, a tentative tectonic history has been assembled by Howard et al. (1974), Strom et al. (1975), Murray et al. (1975), and Dzurisin (1978). The major elements appear to be (1) differentiation of crust, mantle, and core; (2) cooling of a volcanic surface during heavy meteorite bombardment; (3) faulting of the lithosphere by despinning and contraction strains; (4) volcanic and isostatic (?) response to a major meteorite impact; and (5) quiescence, with minor meteorite erosion. Because much of this history is shared by the Moon, we here focus on the unique features of Mercury, which are its volcanic highlands, the Caloris Basin, and strong compression of the lithosphere.

On the Moon, the early crust of the highlands was cratered to saturation with little sign of continuing volcanism. Mercury's intercrater plains, which are the oldest surviving provinces, contain numerous "ghost craters" that were overlapped by flows of almost equal antiquity (Guest et al., 1979). The suggestion is that the early thin crust was denser than the lavas, allowing it to be isostatically depressed beneath them and recycled into the interior. For this to be true, any density difference resulting from the different compositions of the crust and the lavas would have to be smaller than that produced by the effect of temperature on density (up to 10%). In other words, the early crust and the lavas had similar compositions. This is the first major difference from the Moon, where a thick feldspar crust grew on the early magma ocean and later controlled the basaltic volcanism coming out of the mantle by blocking its ascent with a lower-density layer. On Mercury the situation was different: Either the lavas were derived by remelting of the crust, or else no crust ever formed.

The latter option is dubious at best. The albedo and color of Mercury's surface today have been interpreted as indicating a feldspar-rich composition (Hapke et al., 1975). Also, if comparative planetology has any merit, Mercury must have also had a magma ocean. (It is heavier, closer to the Sun, and has a large core.) So it would be hard to explain why feldspars never rose to choke the surface of Mercury. (Kaula, in Chapter 4, suggests that Mercury condensed with too little silica to form feldspars at all, which is a possible rebuttal to this argument.) However, it is far simpler to assume that the early lavas on Mercury came from remelting of even earlier crust, due to heating from strong convection in the mantle below. The overturn of the crust by volcanism would then be a form of layered convection early in the planet's history. When the lithosphere thickened to include the whole crust, this phase came to an end.

Toward the end of the heavy meteorite bombardment, Mercury suffered an especially large impact by a planetesimal, which created the huge impact scar of 1300-km diameter known as the Caloris Basin. Knobby and lineated terrains antipodal to Caloris are attributed to focusing of shock waves created by this impact (Strom, 1979). From the previous discussion (and drawing analogies to the Moon), we would expect such a large impact to make a hole through the entire thickness of the lithosphere. Indeed, the interior of Caloris today is largely a smooth plain interpreted as lava flows. These flows subsided as they cooled, and the decreased surface area of the outer skin (as it increased its radius of curvature within relatively rigid crater walls) led to wrinkling of the surface (Hapke, et al., 1975).

Because no late lava flows continued to erupt and smooth and level this surface (as on the Moon), it may be that the impact postdated the solidification of any global magma layer, and that the flooding occurred by the mechanism of pressure-relief melting discussed above. Unfortunately, the lack of late flows means that no uncompensated load should have formed within Caloris as it cooled; thus it cannot be used to test for the presence of isostatic-adjustment tectonics.

After the formation of a lithosphere and the Caloris impact, Mercury was deformed and faulted on a grand scale. Very spectacular "lobate scarps" up to 3 km tall and 500 km long are known to mark thrust faults because they cut through and shorten older craters (Strom et al., 1975). These scarps occur at all latitudes and, although sinuous, generally have trends within 45° of the North-South principal trend (Dzurisin, 1978). There may also be a preferred orientation of lineaments with trends N60°W and N60°E all around the planet (Dzurisin, 1978), which, because of their straightness and conjugate pattern, would probably be strike-slip faults. Both features could have been formed simultaneously in a stress field where east-west compression was the greatest and the vertical compression was the least. This would be evidence for despinning of Mercury (which now rotates only three times in every two orbits) combined with nonuniform contraction.

However, Strom (1979) argues that the lineations (if real) are older than the lobate scarps. If there were a clear division of the ages, this would invalidate the despinning model. But a combined model, with early despinning accompanied by nonuniform contraction (which continued after despinning) would be consistent with all our present knowledge. The necessary radius change has been estimated by Strom et al. (1975) to be 1 to 2 km. This could be achieved thermally by cooling the whole interior about 40° to 160°C, relative to any cooling of the lithosphere. Or the same contraction could have occurred in response to a smaller temperature drop, if the liquid-solid boundary in Mercury's large core migrated outward by some 10 to 20 km (Solomon and Chaiken 1976). Whatever combination of these two effects is responsible, the implied cooling is quite modest and does *not* preclude a substantial asthenosphere today.

8-4-3 Mars

In the exploration of the solar system, Mars gave us the first exciting look at another planet with active tectonics. Today the activity is modest, probably confined to volcanism, landsliding, and faulting in the Tharsis region. Yet there is reason to think that Mars underwent a more active early stage of horizontal lithospheric movements and "continental drift."

One of the remarkable facts about Mars is its division into two hemispheres of different elevation and age. The southern and higher terrain is the older, on the basis of crater-counts—although not so old as the lunar highlands, if meteorite fluxes were equal. It has not yet been sampled by landers, as both *Viking* missions went to the safer northern plains. Still, we can determine two important things about the south by examining the large, circular impact basins known as planitae (Hellas, Argyre, and several more subdued planitae). First, the large basin relief of 2 km to 6 km (Christensen, 1975) suggests that the impacts blasted holes through an early crust that was much less dense than the mantle-derived lavas that quickly flooded the floors. If this crust did form in the same way as the lunar highlands, then it was probably once distributed uniformly over the whole planet, in that unrecorded (but hypothesized) time when the lithosphere was thinner than the crust. Second, we can infer from the ring-fault structures around basins as small as 120 km in diameter (Guest et al., 1979) that these craters penetrated through the lithosphere (Melosh and McKinnon, 1978), which therefore could not have been more than 60 km thick during these late impacts.

After most of the basins had formed, when the lithosphere was perhaps Earth-like in thickness, something happened to remove this crust from the northern half of the planet. Today the northern plains (Vastitas Borealis) are 5 km lower, less heavily cratered

(Condit, 1978), and appear to be floored by lava flows (Scott, 1978) of basaltic appearance (Binder et al., 1977) and density (Moore et al., 1977). It is inconceivable that these lavas are so much denser than the Martian mantle that they have depressed the early crust 5 to 10 km down and out of sight. It is equally unlikely that the early crust of the north was eroded away, because there is no possible place for storage of such a volume (2×10^9 km^3) of light material. Therefore, it must have moved *laterally* to accumulate in the south.

The most plausible cause for this is that Mars had an early period of homogeneous convection involving the surface while the lithosphere was still thin. This would have produced drifting "continents" of low-density crust, whose trailing margins would preserve the normal faults and grabens formed when they rifted away from another continental mass. In fact, the "fretted terrain" at the highlands–north plains boundary looks very similar to seismic-reflection profiles of the Red Sea and Atlantic margins on Earth (after water and sediments are stripped away). Where continents coverged, such structures would have been crushed, and younger mountain chains uplifted, as the buoyant continents "choked" the downwelling zones and forced the convection to reorganize. Despite billions of years of cratering, erosion, and isostatic adjustment, the sympathetic eye can still find lineations in the topography of the southern highlands that may represent these old continental sutures (e.g., west and north of Hellas, north-east-trending; or west of Chryse, east-west-trending). Of course, detecting subtle lineations on Mars is an old, but not necessarily venerable, tradition!

In this view, the north plains are like seafloor on Earth, created by partial melting and volcanism at sites of upwelling. The reasons no topographic rises are visible today are that the system has been stopped for over 2 Gy and the lithosphere thickness has equalized. What stopped the overturn was apparently the collision of all the original crust into a southern "supercontinent," like Pangaea on Earth, which formed at 0.2 Gy. The present assembly is not necessarily the only or the first one; it would be the particular one that occurred after the lithosphere reached a critical thickness that prevented convection, once stopped, from reinitiating itself.

To gauge the lithosphere today, we can use the great active (or recently extinct) volcano Olympus Mons. Dividing its height of 25 km by the maximum plausible lava-rock density contrast of 10 percent, we conclude that the lava column must be confined in a strong lithosphere to at least 250 km in order to erupt. Such a thick lithosphere would have tremendous strength and could easily account for the confinement of planetary convection within the interior today. In fact, some have speculated that it was strong enough to support Olympus, the three volcanoes of the Tharsis Ridge, and the vast elevated terrain surrounding them for well over 1 Gy, while developing only minor surficial faults.

The great mound of Tharsis Montes corresponds to about 90° in longitude, and the mound rises above its surroundings to heights variously estimated as 9 km or 4 km (Downs et al., 1982). Sitting atop it are the three volcanoes Arsia, Pavonis, and Ascraeus Mons, which rise an additional 15 km above the mound. (Their collective mass is dwarfed by the excess material of Tharsis Montes.) Most early interpretations of this topography assumed that the surface of Tharsis is only thinly covered with lava flows and that the original flat surface of Mars has been warped upward by internal convection. Now, however, evidence is accumulating that volcanism in the Tharsis region has continued over most of the planet's history (Plescia and Saunders, 1982), so that the mound may be formed from a very thick accumulation of lava flows (Solomon and Head, 1982b; Willeman and Turcotte, 1982) and the original surface may actually be warped down beneath this load.

The support of the Tharsis bulge is the great tectonic problem of Mars, and (as the Introduction in this chapter warned) both the amount of load and the response are controversial. Figure 8-4 shows the major possibilities schematically in cross section. A valuable constraint is the gravity field, which shows a broad positive anomaly above Tharsis, but with only about 30 percent of the size it should have in the totally uncompensated case (Christensen and Balmino, 1979). This allows one to choose any model between the extremes of 70% shallow (crustal) compensation (with considerable stress in the lithosphere) and total compensation distributed deeply (from near the surface to a depth of 300 km to 400 km). Crustal compensation could be created simply by volcanic thickening of preexisting crust; for plausible density contrasts, some 20 km to 50 km of thickening would be required. The cause for deep compensation could be replacement of colder, denser lithosphere under Tharsis and Olympus by broad instrusions of hotter, lighter asthenosphere with the same composition (see Fig. 8-4). Such a "hot spot" could only be maintained by steady convective upwelling of a great plume or plumes beneath the region, which naturally would also explain why melting and volcanism are so severely restricted today. (Theoretical studies have not yet advanced far enough to predict whether such plumes under thick motionless lithosphere are stable for billions of years; however we can predict that if such a plume did move, the abandoned thermal dome would decay down to average elevation with a very long characteristic time of 1 to 2 Gy. There is no clear sign of this on Mars.)

One additional clue to the solution is the observation of radial rilles all around Tharsis, extending out at least 60° from its center. These rilles are thought to be downfaulted blocks caused by normal faulting and crustal stretching in the direction circum-

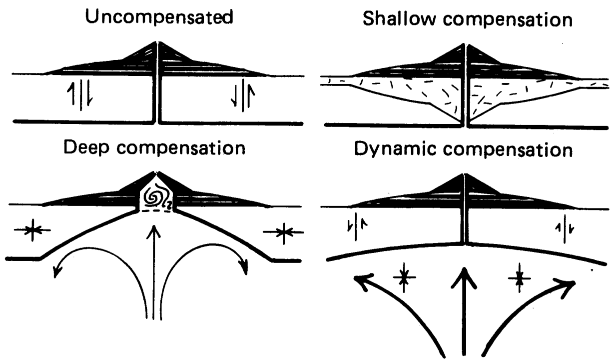

Fig. 8-4 Four hypotheses for the support of the Tharsis bulge and superjacent volcanoes on Mars. Heavy line indicates lithosphere-asthenosphere boundary; crust has random dashes. Paired arrows indicate stress, and single arrows show convective flow. Gravity and fault patterns seem to favor a mixed model, with part of the load uncompensated, and part compensated deeply by lithospheric thinning. The role of shallow compensation is uncertain.

ferential to Tharis (Carr, 1974). Their close relationship to the volcanic loading is shown by a mutual overlapping of rilles and flows extending far back in Martian history (Plescia and Saunders, 1982). Also, the great rift of Valles Marinerus, which is probably just the longest of these radial grabens, displays an intercutting pattern of scarps, landslides, and craters, which implies that it is still active (Blasius et al., 1977). A number of investigators (Banerdt et al., 1982; Solomon and Head, 1982b; Willeman and Turcotte, 1982), have attempted to compute models of a thick lithosphere subject to various vertical loads that will match these observations as well as Martian topography and gravity data. One clear consensus is that the uncompensated part of the Tharsis load is laterally supported by the strength of the surrounding lithosphere rather than by very fast (and viscous) convection in the asthenosphere. This is necessary to explain the great extent of the rilles and the activity of the Valles Marineris. Less obvious is whether the partial compensation is shallow (thickened crust), deep (heated and thinned lithosphere), or both. However, the simple fact of continuing volcanism implies a great and concentrated heat source, probably an upwelling mantle plume. Therefore deep compensation in the form of lithospheric thinning under Tharsis must play some role. Banerdt et al. (1982) have shown mathematically that some component of deep compensation is helpful in explaining the extension of the rilles into very small radii around Tharsis; in a pure bending model these would be suppressed at radii of less than 30°. Another factor that could be important is superposed nonuniform contraction, with the exterior lithosphere cooling more than the internal asthenosphere. This would tend to encourage normal faulting all over the surface of Mars (Solomon and Chaiken, 1976), reinforcing the effect of the weight of Tharsis.

8-4-4 Earth

The Earth today has a relatively thin lithosphere. Under the oceans it thickens as the square root of its age, from zero at spreading centers to 30 km in the oldest parts of the Pacific (Goetze and Evans, 1979). In continents, the situation is more complex. Earthquake locations and extrapolated creep laws both show a lithosphere-asthenosphere transition *within* the crust at 10- to 25-km depths (Sibson, 1982). However, the uppermost mantle just below the Moho is also a layer of great strength (in regions of low to average heat flow), because mantle olivine has a higher creep-softening temperature than crustal quartz and feldspars. This stiff upper mantle restricts the flow of the soft lower crust enough to maintain the integrity of continents. However, it must be remembered that when orogenies occur within continents, the upper crust and upper mantle generally decouple and deform with quite different styles (e.g., Bird and Rosenstock, 1984).

Plate tectonics as the dominant style of the Earth has been abundantly documented elsewhere (Le Pichon et al., 1976). Likewise, the existence of upwelling plumes beneath Hawaii, Iceland, Yellowstone, and a dozen other localities is no longer controversial. Together, these two mechanisms (and the associated isostatic adjustments) explain almost all of the topography, geology, and volcanic deposits forming today. A few remaining anomalies require other mechanisms from the theoretical list; the author has suggested delamination as the cause for migrating waves of uplift and volcanism within the western United States (Bird, 1979) and put forward a variant of layered convection with "invisible" subsurface mantle downwelling as an explanation for the seismic velocity anomaly beneath the Transverse Ranges of California (Bird and Rosenstock, 1984). Despinning of the Earth is well documented (Munk and McDonald, 1960), but associated faulting has never been confidently identified amid the welter of larger deformations produced more rapidly by plate tectonics. Similarly, the fragmentation of Earth's lithosphere into many changing plates makes it unlikely that nonuniform contraction effects will ever be identified, although it has almost certainly occurred (Schubert et al., 1980). This mechanism was once the last resort of theoreticians (e.g., Bucher, 1933, p. 123) for all unexplained orogenies, before the understanding of plate tectonics. This should serve as a caution that our present synthesis may also be subject to drastic revision.

The present synthesis of Earth tectonics must also be qualified by saying that it may only apply to the last billion years; this is the greatest age of rocks obviously formed by sea-floor spreading or subduction. The student of early Earth history is in some ways much poorer in data than the student of the Moon: although all the desired rock samples are available, there is absolutely no information about the topography, gravity, heat flow, seismicity, crustal thickness, planetary radius, or rotation rate prevailing at the time they formed. The rocks of the Archean Era (i.e., those from the first half of Earth history) available for study show an association of "greenstone" belts (metamorphosed basalt flows and sandstones) and grey tonalitic gneisses with younger granite intrusions. These bodies of rock have small dimensions (5 to 50 km) and ambiguous structural relations (in West Australia the greenstones surround the gneisses; in the Canadian shield, vice versa). Stratigraphic relations show that the gneisses rose above sea level and the greenstone belts formed below (Kummel, 1961). This observation shows that sea level was high enough to cover some, but not all of the area of continental crust (as it is today); therefore we can infer that the ratio of Archean crust to ocean was similar to the present one.

With regard to the global tectonic regime, there are two other useful constraints: The surface temperature has remained in the general area of 0°C to 100°C during the last 3 Gy (to permit the survival of algal life), and the heat flow has declined by a factor of at least three in the same time due to radioactive decay and planetary cooling (Schubert et al., 1980). If we apply the scaling laws of convection theory, the joint implication of these facts is that at 3 Gy ago, the mantle was about 200°C hotter and the speed of convection about five times as great.

The effect that this would have on surface tectonics depends on how much continental crust and ocean had already formed, but this is not known. One extreme possibility is that all continental crust differentiated immediately and has merely been reworked and remelted ever since. In that case, the higher Archean heat flow would have led to excellent crust-mantle decoupling in the soft lower crust and the development of layered convection. The granite bodies would mark sites of upwelling in this model, while the greenstone belts would represent crustal downwelling sites that continuously engulfed and recycled sediments. (No obvious source for the basalt flows is implied by this model.) Conversely, if the amounts of Archean crust and ocean were small, then erosion and deposition would have redistributed the crust into thin patches. The temperature at the base of the crust would have been low despite the high heat flow, and the crust would have been firmly attached to the convecting mantle. We would then interpret the granitic blocks as microcontinents assembled and sutured together in the downwelling zones of a homogeneous or plate-tectonic convection system arrested by buoyancy, and interpret the greenstone belts as remnants of lost oceanic crust that once separated them.

We have no record at all of events on Earth with ages exceeding 3.8 Gy, because no rocks have survived unaltered. Presumably the Earth is at least as old as the Moon, and a significant period of time passed when crust was either not differentiated or not preserved here. The argument that it never formed is weakened by evidence of early crustal layers on the three smaller (and cooler) planets we have already examined; therefore destruction or reheating is indicated. Under the layered-convection model proposed above, it could be argued that it took 1 Gy for the top of the crust to form a lithosphere strong enough to resist recycling through the crustal convecting layer, with consequent resetting of radiometric ages. Or, in the microcontinent model, it would be argued that the size and number of crustal blocks had to grow to a critical size before their buoyancy was great enough to prevent their being dragged down into the mantle as a part of the circulating noncontinental lithosphere. A decision between these models will probably have to be made on the basis of geochemical arguments constraining the history of the total mass of differentiated crust. Preliminary indications from neodymium-isotope ratios (DePaolo, 1981) are that crustal fractionation has been gradual (see Chapter 5 by DePaolo); this favors the microcontinent model of homogeneous or plate-tectonic style.

The homogeneous convection style, in which downwelling is symmetrical, is a better choice for the Earth's first eon, because it avoids two objections that have been raised to early plate tectonics. These are the absences of (1) Archean "blueschists," to indicate subduction, and (2) basalts intruding older basalts, to indicate sea-floor spreading (Kröner, 1979). Blueschists of the recent plate-tectonic age form when sediments are caught in a zone of recirculating flow between one subducting and one stationary plate (Cloos, 1982) where the ratio of temperature to pressure is especially low. But sediments caught between two plates that were both sinking would not have been returned to the surface (except in an adjacent upwelling zone that would be too hot to form blueschist). Ocean floor, on the other hand, may have been formed in the Archean Era, but not preserved. Even today the uplift of ocean-floor basalts onto continents is a rare event and is probably caused by abortive plate-tectonic subduction of a buoyant continent beneath an oceanic arc. Without asymmetrical subduction, these rocks would never have been lifted high enough to be visible on land today.

As the Earth gradually cooled and its lithosphere thickened, the forces required to bend downwelling lithosphere through a tight curve down into a trench would have increased roughly at the cube of its thickness (Goetze and Evans, 1979). (Specifically, the bending moment per unit length of trench is proportional to the cube of thickness.) On the other hand, the convective density anomalies that gravity acts upon to create these bending moments would have increased more slowly as the square of lithosphere thickness. This means that the convecting system would have been thrown out of balance by the thickening of the lithosphere, and bending at trenches would come to consume more and more of the energy available. There would have been a decrease in the stability of the old homogeneous convection with respect to plate tectonics, in which one plate is not deformed and the other bends very gently through a large arc. The resulting change of style apparently occurred around 1 Gy ago. The new style of plate tectonics would not have been possible without an ocean, because it is the pore water dragged into the fault between the converging plates that reduces friction and makes the sliding possible (Bird, 1978). If Earth had been as dry as Mars, its horizontal tectonics might also have come to an end eons ago!

In comparing Earth to other planets, it is essential to realize how much our hydrosphere controls the phenomena we regard as normal. Water not only lubricates subduction, but is directly responsible for the generation of magmas in downwelling zones, despite their anomalously low temperatures. Many believe that the present continental crust is not primary, but was formed by differentiation of basaltic magmas in these arcs (Wyllie, 1982) and by further sorting during weathering and sedimentation. Likewise, "oceanic" crust (i.e., that found between the granite "continents") could be different on a dry Earth, because the present mantle temperature is probably too low for convection alone to produce any melting in the absence of volatiles. The major physiographic characteristic of the Earth, the neat separation of oceanic and continental crust into two levels, is also a physical result of the ocean. Without it, the upper level would vanish as erosion and isostatic rebound distributed any crust more uniformly around the planet. In short, our planet might have a "frozen" global lithosphere, no horizontal tectonics, no continents or ocean basins, but only a monotonous thin covering of sand on an irregular ancient surface, occasionally relieved by plume volcanism and vertical tectonics. This exercise in imagination is a very good introduction to our neighboring planet and closest analogue.

8-4-5 Venus

Venus is the most poorly known of the five planets because of its thick cloud cover, and thus the interpretation of its tectonics is still very primitive. Four successful *Venera* soft landings were made east of the elevated Beta Regio, but the chemical analyses returned are inconclusive; either a granitic or a basaltic crustal composition can be inferred. (Without knowing whether the rocks are igneous, metamorphic, or sedimentary, we can infer at least local differentiation and volcanism from the fact that either rock composition would yield the wrong total mass if extrapolated throughout the planet.) Next in terms of the resolution of view obtained are the Earth-based (Arecibo) radar observations reported by Campbell and Burns (1980). Despite coverage of almost half the surface at 5-km to 20-km resolutions, no evidence for a grid of linear faults has yet been found. Thus there is no reason to think that despinning and nonuniform contraction have been important in shaping the present surface of Venus. Isostatic readjustment cannot be properly studied without detailed topographic data over unambiguous volcanoes or impact craters, but we do have data on global topography from satellite altimetry (Masursky et al., 1980), and global gravity from satellite orbits (Sjogren et al., 1980), which can be compared. Although the two are correlated, the gravity anomalies over broad features are only 15 to 35 percent of the attraction of the topography (Sjogren et al., 1980), so at least partial isostatic compensation is required. Phillips et al. (1981) have made a strong case that the remaining gravity anomalies probably result from deep compensation at 85- to 150-km depths, rather than from uncompensated topography supported by a strong lithosphere.

The crucial remaining question is whether the surface of Venus is a global lithosphere, or whether it convects. Kaula and Phillips (1981) have searched the topographic data for any elevated linear features corresponding to plate-tectonic spreading centers; the result was that the "best candidates" have inconsistent elevations, a disorganized and disconnected map pattern, and a shape different from that predicted by convection theory. They took this to mean that homogeneous and plate-tectonic convection are absent, or at best they are slow and unimportant in transmitting the total heat flow. Additional arguments for a global-lithosphere model are based on some 50 quasi-circular features in the radar images, with diameters of 20 to 1200 km (Campbell and Burns, 1980). If these are interpreted as impact craters, and if meteorite-influx histories of Venus and the Moon are comparable, this implies that at least part of the crust on Venus is over 1 Gy old and cannot be involved in convection (Masursky et al., 1980). On this basis, Phillips et al. (1981) have presented a Venus model in which plume convection dominates. In their view, hot upwelling plumes lie beneath each of the major highlands (Ishtar Terra, Aphrodite Terra, Beta Regio) and provide both a magma source for construction of continental-sized volcano complexes, and the thermal buoyancy needed to isostatically support them. A difficulty with this model is that it cannot explain the deep curvilinear troughs (Artemis, Dali, Devana, and Diana Chasmae) adjacent to these highlands; they must be attributed to an earlier, arrested stage of surface convection. It is also difficult to explain why these have not been filled with sediment, considering that dense Venusian winds should be able to transport sand and gravel up to 1 cm in diameter (Warner, 1982).

It must be admitted that the global-lithosphere or "Martian" model *is* consistent with known rock mechanics, even though Venus is much hotter than Mars. Consider the steep slopes at the edge of the terrae, such as the "Vesta Rupes" southwest of Ishtar, which slopes down 3 km in 350 km. Using approximate values of gravity and density, we find the shear stress beneath this slope must increase with depth at a rate of $0.24\ \mathrm{MPa} \cdot \mathrm{km}^{-1}$ down to about 200-km depths. Meanwhile, temperature increases from the surface value of 460°C at a gradient of roughly $15°\mathrm{C} \cdot \mathrm{km}^{-1}$ (Phillips et al., 1981). Combining these conditions, we find that the strain rate from dislocation creep becomes significant ($10^{-8}\ \mathrm{y}^{-1}$) at only 28-km depth with an olivine or gabbroic flow-law, or at only 9 km with a granitic flow-law (Tullis and Yund, 1981). In other words, a thin lithosphere must act alone to support the horizontal normal stress that keeps Ishtar from spreading and collapsing. This stress (approximately = 3 km height $\times$ $8.9\ \mathrm{m} \cdot \mathrm{s}^{-2}$ gravity $\times$ $3000\ \mathrm{kg} \cdot \mathrm{m}^{-3}$ density $\times$ 100 km depth-to-compensation $\div$ lithosphere thickness) becomes 280 to 900 MPa, depending on the composition at these depths. The lower figure is consistent with the limits imposed by frictional sliding, and the upper figure is not. That is, Ishtar Terra *could* be a static feature *if* it contains less than 9 km of granitic crust. A purely gabbroic crust is quite possible, considering the speculations of the previous section on the control of planetary differentiation by the presence or absence of water. Perhaps the 3-km height of Ishtar is determined by the fact that gabbroic crustal roots (with a 10 percent density contrast) cannot extend more than 30 km down and still remain part of the lithosphere.

Although this "Martian model" is self-consistent, the linear troughs are a perplexing puzzle and lead us to entertain alternatives. Perhaps some of the circular features are not impact craters, for they fail to show the expected central depression (Masursky et al., 1980) or circular scarps (Campbell and Burns, 1980). (Solomon et al. (1982) have pointed out that significant viscous relaxation of crater forms is to be expected on Venus; this type of argument can be used either to excuse the peculiar shape of these "craters" or, conversely, to argue that they must be very young.) On a planet that is so hot, we can equally well speculate that these are volcanic features, such as collapsed calderas or eroded intrusion domes on Earth. Or, with more data we may find that these *are* ancient impact craters, but that they are confined to areas of ancient surface surrounded by mobile belts. In that case crustal convection is possible (especially with a granitic composition). Solomon and Head (1982a) have already shown that all the other arguments against convection are inconclusive.

It was already mentioned above that layered convection can produce any finite ratio of gravity to topography, including the one observed on Venus. Likewise, a wide variety of topographic forms is possible. The spreading centers in the crustal layer would necessarily be the highest points, but could have different absolute heights depending on the thickness of the mantle boundary layers they overlay. Furthermore, the colder, more brittle upper crust could fault and deform in a chaotic fashion in mantle-convergence-crustal-divergence zones, where mantle heat flow is low (see Fig. 8-5). Conversely, the thickening of crustal upper-boundary layers could be arrested or reversed by high heat flow near mantle-divergence zones, producing more ductile flow and gentler topography. Additional complications could result if the crust was insufficient to cover the whole planet: Where mantle upwelling swept aside the crust, the topography could be W-shaped in profile (Tellus Regio?), and where mantle convergence was unobscured by the crust, there would be symmetrical outer rises about a central trough (Dali Chasma?). (Alternatively, the chasms could indicate grabens formed by crustal extension above mantle downwelling.) Such steep asymmetrical scarps as the Vesta Rupes could be formed by intracrustal thrust faults, like the Himalayan front on Earth, that maintain steep slopes by constantly sweeping up a soft viscous crust.

These widely divergent possibilities for Venus could be constrained by just a little more data. A clear view of the surface (which will soon be obtained by synthetic-aperture radar imaging from orbit) might clear up the crucial questions of origin and distribution

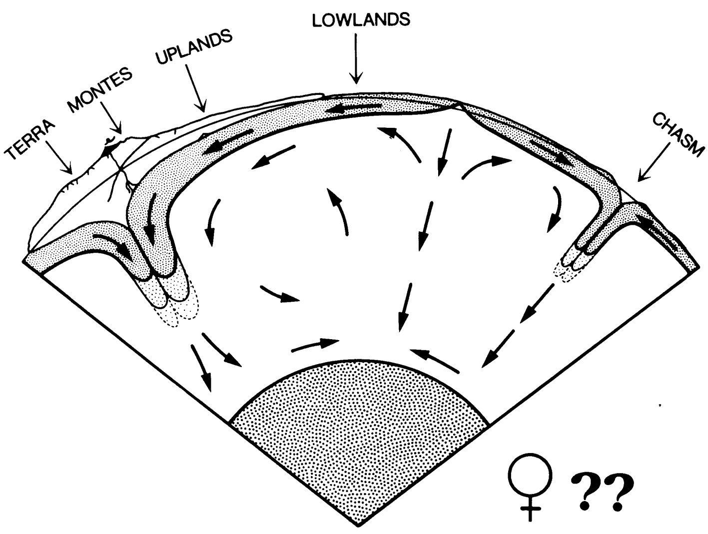

Fig. 8-5 Cartoon of a layered-convection model for Venus, with considerable vertical exaggeration. Crust and asthenosphere are white, and mantle lithosphere is shaded. This model is only applicable if Venus has a large amount of silicic crust, which is mobile at near-surface temperatures. It predicts large amounts of young surface, hence could be invalidated if circular features on Venus prove to be ancient impact craters and are uniformly distributed.

of the small circular features. Wider sampling of crustal composition by landers would be very helpful, as would technological innovations allowing a heat flow experiment to survive on the surface. The exploration of Venus (see Chapter 17 by Kivelson) remains one of the most exciting and accessible problems in space.

8-5 CONCLUSIONS

A detailed survey of the inner five terrestrial planets for tectonic styles suggested by theory reveals possible examples of every type (see Fig. 8-3). Each planet has had volcanism at least in its early history; on the Moon this may have been a passive response to crater excavation, but on the other planets it seems to indicate a retention of volatiles that depressed the melting point enough to match the temperatures in upwelling convection.

On Mercury, the lavas may have been remelted crust, recirculating in the upper level of an early layered-convection system. Each planet also shows evidence of at least partial isostatic compensation of larger and older features, confirming that the theoretical concept of a thickening lithosphere over a convecting interior is universally valid for stony planets. In each case, the gradual loss of primordial heat of accretion and differentiation and the decay of radioactive elements have decreased the planetary heat flow and thickened the lithosphere, leading to a decline of tectonics. Mercury and the Moon have lithospheres that were never broken by internal convection, but only by the small strains of isostatic adjustment, nonuniform contraction, and perhaps despinning. Mars probably had homogeneous convection in its early history, which collected the ancient (granitic?) crust in the South and formed waterless basaltic "oceans" in the North; but today horizontal motions have stopped, and even the plumes that rise to the surface at Tharsis do not wander. The Earth has probably differentiated its oceans and continental crust gradually over time. The simultaneous thickening of the crust and the lithosphere forced two changes of convective style, from steady-state homogeneous convection to "jerky" homogeneous convection that was locally arrested by buoyant continents, and then from homogeneous convection to the asymmetric plate-tectonic style. It is likely that surface convection on Earth today is only possible because of water lubrication of subduction zones. Venus presents a dilemma, because its greenhouse effect has boiled the water off a surface that would otherwise be much like the Earth's. If this happened early, then Venus probably has only basaltic crust of great strength and high melting point, so that its tectonics are like those of Mars despite a 500°C difference in surface temperature. However, if granitic crust formed before the water was lost, it would be highly mobile today. The continent-sized *terrae* would then have to be dynamic features, probably overturning piles of buoyant crust swept together by mantle convergence in the lower level of a layered-convection system.

8-6 ACKNOWLEDGMENTS

As a newcomer to the field of extraterrestrial geology, I must stress that none of the data and few of the concepts in the review are original. The authors of the references cited (which are only representative and not exhaustive) were the original developers of the

models presented here, excepting only delamination and layered convection. Very valuable discussions that helped to form my ideas were held with W. M. Kaula about Venus, with D. G. Sandwell and P. R. Mullen about Mars, and with W. G. Ernst and S. Peacock concerning the early history of the Earth. W. M. Kaula and S. C. Solomon each reviewed the manuscript and suggested a number of significant improvements.

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