Tectonic Evolution of the Terrestrial Planets

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The surface features of a planet provide important clues to both internal and external geological processes. For example, we can read much about the state and thermal evolution of the interior of Earth from patterns of volcanism and near-surface deformation through time. The major surface features of the other terrestrial planets have been revealed by images returned from U.S. and Soviet space probes over the past 15 years. Study of the surface features of these planets has led to an improved understanding of the processes that shaped their surface and interior evolution and to the recognition that, in contrast with Earth, much of the evolution of the smaller terrestrial planets was concentrated into the first half of the history of the solar system (*1*). This article discusses the tectonic style of each terrestrial planet as inferred from its surface features, compares these styles with models of planetary thermal history, and reviews the major factors responsible for, and the processes governing, planetary tectonic evolution.

Summary. The style and evolution of tectonics on the terrestrial planets differ substantially. The style is related to the thickness of the lithosphere and to whether the lithosphere is divided into distinct, mobile plates that can be recycled into the mantle, as on Earth, or is a single spherical shell, as on the moon, Mars, and Mercury. The evolution of a planetary lithosphere and the development of plate tectonics appear to be influenced by several factors, including planetary size, chemistry, and external and internal heat sources. Vertical tectonic movement due to lithospheric loading or uplift is similar on all of the terrestrial planets and is controlled by the local thickness and rheology of the lithosphere. The surface of Venus, although known only at low resolution, displays features both similar to those on Earth (mountain belts, high plateaus) and similar to those on the smaller planets (possible impact basins). Improved understanding of the tectonic evolution of Venus will permit an evaluation of the relative roles of planetary size and chemistry in determining evolutionary style.

Three major influences on the evolution of planetary bodies are known, and their detailed effects on specific planets is a subject of current research. The first of these is planetary radius. The size of a planetary body affects its subsequent thermal evolution through the ratio of heat radiated at the surface to heat generated in the planetary volume (*2*) and because of the strong dependence of the Rayleigh number, a measure of the convective vigor of a planetary mantle, on the radial length scale (*3*). The second major influence is planetary chemistry, including particularly the iron/silicate ratio, the abundance of radiogenic heat sources, and the abundance of volatiles. One important factor for the bulk chemistry of the planets is thought to be distance from the sun, because of the decrease in temperature and pressure outward from the center of the early solar nebula during the condensation of major elements and compounds (*4*). The third major influence is the budget of internal and external energy sources during the thermal evolution of the planet. These sources may include energy associated with accretion and subsequent bombardment, differentiation, tidal interaction, radioactivity, and electromagnetic heating (*5*).

The link between the internal thermal evolution of a planet and its surface geologic history is the lithosphere, the outermost layer of a planetary body capable of maintaining elastic stress differences over geological time. The base of the lithosphere is marked by a change in the rheological properties of the material, and its position is sensitive to temperature, composition, and strain rate (*6*). The lithosphere should be distinguished from the crust, which is defined on the basis of a compositional difference from the underlying mantle. We use the term tectonics to denote the large-scale deformation of the lithosphere in response to stress. Such deformation generally is accompanied by faulting at places on the planetary surface where stresses exceed elastic strength. There are many sources of stress in planetary lithospheres. They include mantle convection, volcanic or sedimentary loading, temperature changes, tides, real or apparent polar motion, tidal despinning, topographic variations, and dynamic loading (impact cratering) (*7*). The style of lithospheric tectonics is closely related to the thermal evolution of the interior of the planet (*8*). On Earth, the existence of the relation is obvious since the creation and destruction of the lithosphere is a significant part of the mechanism by which heat is transferred out of the mantle. Tectonics and thermal history are also closely related on planets that lack laterally mobile lithospheric plates (the moon, Mars, and Mercury). Global thermal expansion or contraction and associated stresses can result in surface faulting and can directly influence the location and abundance of surface volcanism.

The Moon

The surface of the moon is dominated by craters and basins produced by meteoroid impact. About 80 percent of the surface is composed of feldspar-rich crustal material (*9*) shaped by major impact processes prior to about 4 billion years ago (*10*). Basaltic lavas were emplaced over the remaining 20 percent of the surface between about 2.5 and 4 billion years ago (Fig. 1) (*11*). These lavas formed plains, the maria, located primarily within ancient impact basins in various states of degradation. Tectonic features, consisting mainly of linear rilles (graben) and mare ridges and arches, are located predominantly in and adjacent to the maria (Fig. 2) (*12*). The intense cratering has obliterated morphological evidence of surface and interior processes in the first few hundred million years of lunar history. The gravitational field of

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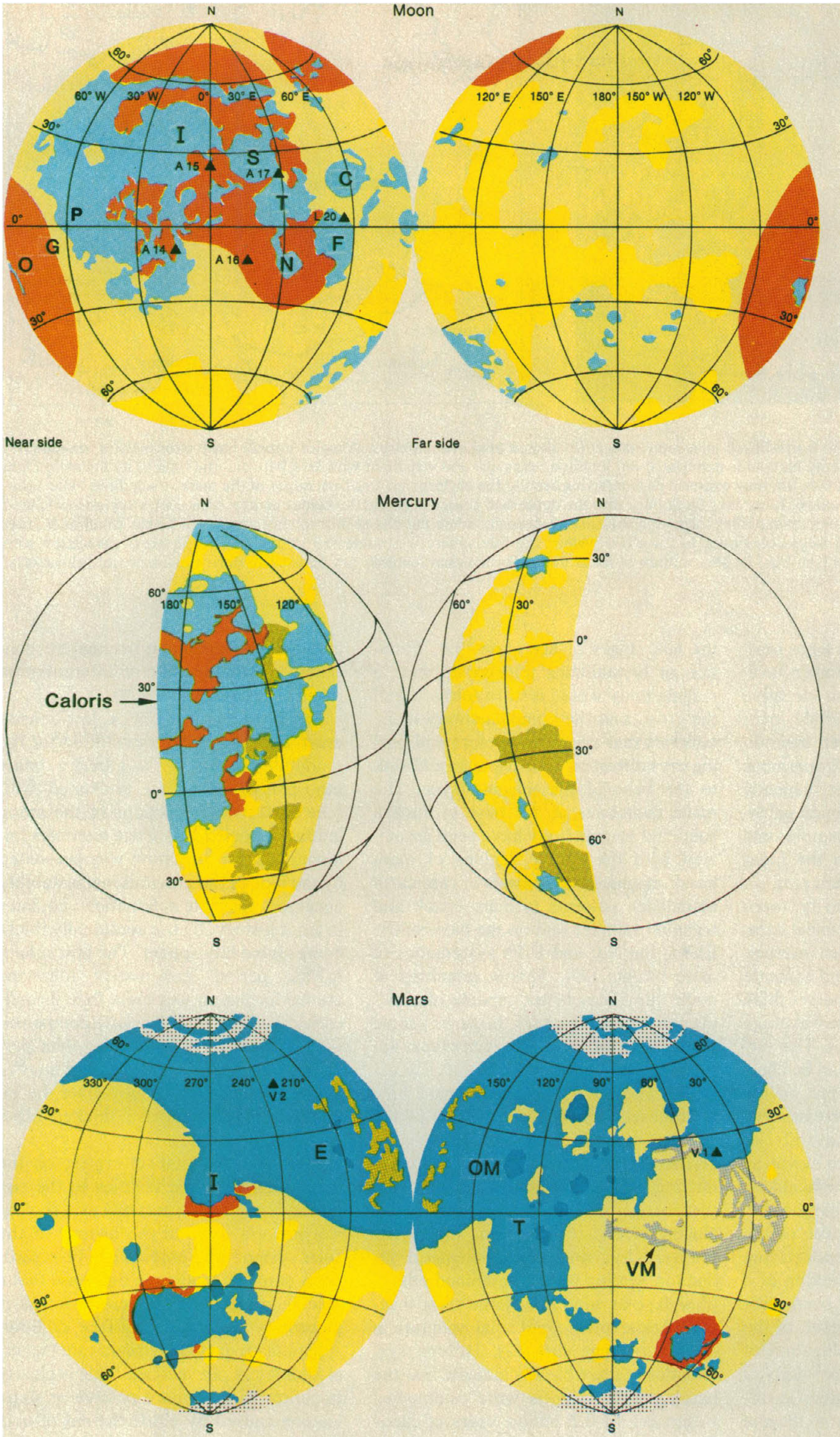

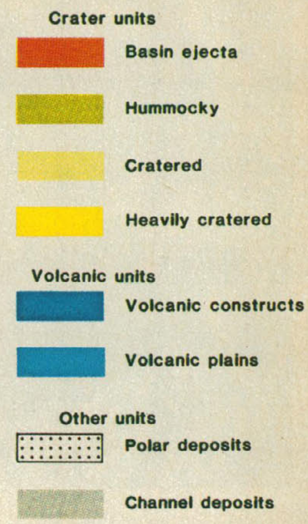

Fig. 1. Geological terrain maps of the moon, Mercury, and Mars (*1*). Lunar maria are indicated by letters: *I*, Imbrium; *S*, Serenitatis; *T*, Tranquillitatis; *F*, Fecunditatis; *C*, Crisium; *O*, Orientale; *P*, Oceanus Procellarum; and *G*, Grimaldi. Only about one-third of Mercury has been viewed at high resolution. Large tracts of smooth lava-like plains occur around the Caloris basin (30°N, 190°). On the moon and Mercury, vast areas of cratered terrain dominate the surface, and volcanic plains form a relatively thin veneer. On Mars, giant shield volcanoes rise from the extensive lava plains (*T*, Tharsis; *OM*, Olympus Mons; *I*, Isidis; *E*, Elysium; *VM*, Valles Marineris; and *V1* and *V2*, Viking landing sites). In general, cratered terrain is dominated by impact craters; heavily cratered terrain, having accumulated more craters, is older. Basin ejecta is material thrown from young, large impact basins. Hummocky terrain may be related to antipodal effects of basin formation. Volcanic plains are basaltic lava flows on the moon and are likely to be comparable extrusive deposits on Mars and Mercury. Volcanic constructs are major topographic accumulations of lava around a source vent, which on Mars often rises 25 km above surrounding terrain. Polar deposits on Mars are layered fragmental deposits and solid water and carbon dioxide; channel deposits appear to have been formed by the flow of liquid water.

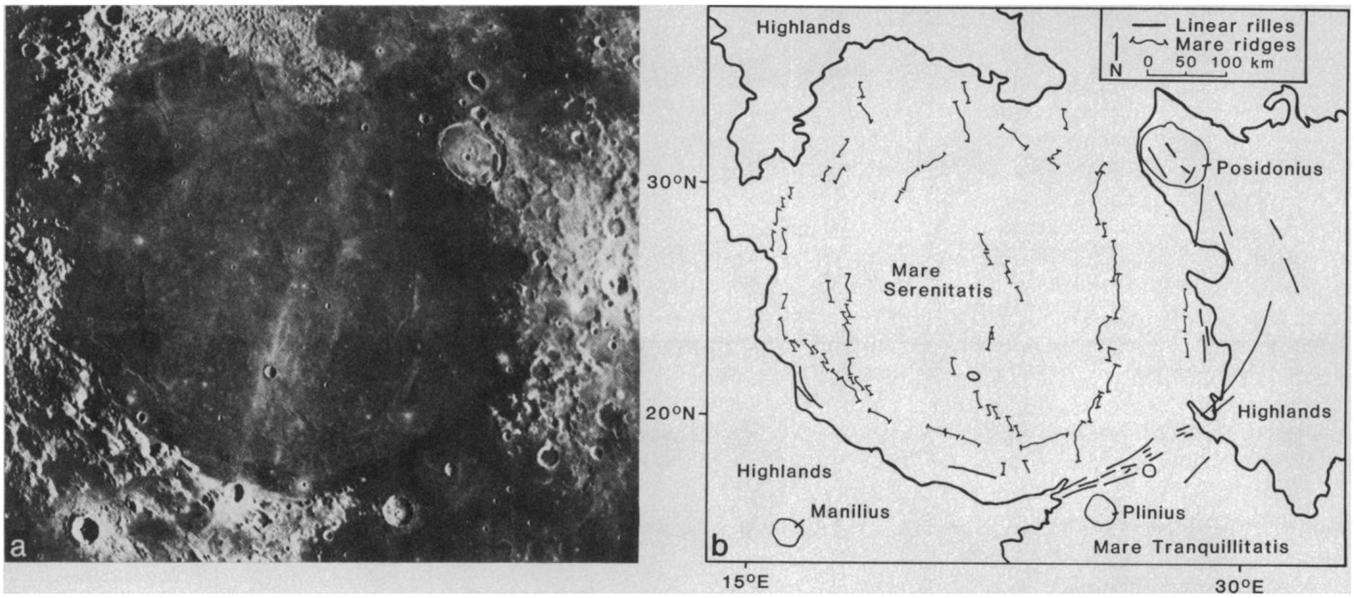

Fig. 2. Lunar Mare Serenitatis. (a) Earth-based telescopic image; (b) sketch map of same area. A major impact basin over 600 km in diameter, Serenitatis was excavated in cratered highlands material about 4 billion years ago and was filled with lava between the time of its formation and about 3 billion years ago. Earliest lava fill, now exposed as a dark ring around the eastern and southern edges of the mare, loaded the lunar lithosphere, causing subsidence and flexure. Later lava fill flooded into the depressed center, forming the lighter central area. Deformation associated with the loading included extensional linear rilles formed in the earliest deposits when the global lithospheric stress was either small or slightly extensional. Compressional mare ridges continued to form later, when the lithosphere was subjected to compressional stress associated with planetary cooling and contraction (see Fig. 4). Distribution of loads and deformational features can be used to derive estimates of lithospheric thickness (see Fig. 5) (*41*).

the moon is characterized by large positive anomalies over the younger lava-filled basins. There are also negative anomalies of smaller magnitude over small unfilled basins and over unmodified impact craters such as Copernicus (*13*). The present dipole lunar magnetic field is insignificant (*14*), although paleomagnetism in returned lunar samples and local magnetic anomalies on the lunar surface indicate an ancient field (*15*).

On the basis of seismic, gravity, topographic, petrological, and chemical data, the moon is known to have been internally differentiated into a crust and a mantle (Fig. 3). The crust is 55 kilometers thick on the central nearside and may average 70 kilometers in thickness (*16*). The surface of the nonmare regions can be characterized as anorthositic gabbro on the basis of chemistry (*9*), and correlation of the surface abundances of several elements (such as Al and Th) with elevation in isostatically compensated highland areas supports the hypothesis that the surface chemistry is representative of the composition of the bulk of the underlying crustal column (*17*). The formation of a thick, feldspar-rich crust has been most commonly explained as the result of the cooling and fractional crystallization of one or more large magma bodies early in lunar history. The magma bodies in different scenarios range from a magma "ocean" several hundred kilometers deep (*18*) to a succession of magma "lakes" associated with large impacts in the late stages of accretion (*19*). Flotation of crystallizing feldspar in such a magma body would lead to formation of the crust, and the complementary mafic minerals that crystallize before and during crystallization of feldspar would sink to the base of the magma layer. The mafic cumulates at the base of such a solidified magma body have been associated with the source regions of mare basalt magmas on the basis of rare earth abundance patterns in mare basalt and highland samples and on the basis of the Rb/Sr, Sm/Nd, and U/Pb systematics of mare basalts (*20*). Simple remelting of mafic cumulates is not capable of yielding mare basalt magmas, however; some more complicated multistage process appears to be required (*21*).

The magma ocean and magma lake hypotheses require early melting of the outer portions of the moon. There is no consensus on the cause of this early melting, but possible candidates include the energy of large impacts during the closing phases of accretion, short-lived radioactivity, and electromagnetic induction heating by the combined effects of high solar wind flux, large solar magnetic fields, and a rapid solar spin rate in a T-Tauri–stage sun (*22*). Isotope systematics of mare basalts indicate that the basalt source regions were established within 0.1 to 0.2 billion years of lunar formation (*23*). Thermal history models of lunar magma oceans which include the effects of gravitational separation of crystals and remaining magma are consistent with completion of differentiation in such a time interval (*24*).

Early formation of a thick, low-density crust led to an interior evolution for the moon dominated by a globally intact lithospheric shell that thickened with time as the outer portions of the moon gradually cooled (*25*). Plate tectonics following crustal formation was prevented by the lack of the gravitational instability necessary for plate subduction and, later in lunar history, by the continually thickening global lithosphere. The lithosphere in the period immediately following crustal formation was quite thin, as indicated by the general isostatic compensation of highland terrain older than perhaps 4 billion years (*26*). Several lines of evidence, however, demonstrate the effects of a substantially thicker lithosphere at later times.

The density of impact craters on the lunar highland crust testifies to the significance of impact cratering as a geological process early in the history of the solar system. The process can be regarded as dynamic loading in the sense that a load (a projectile of a certain mass) is placed on a planetary surface or lithosphere at an extremely rapid rate (several kilometers per second). This is clearly in contrast to the more passive or static loading caused by emplacement of lava flows over thousands or millions of years. We consider many aspects of impact cratering to be tectonic processes because they represent large-scale deformation of the lithosphere in response to stress. Most of the kinetic energy of the impact (~ 70 to 85 percent) is converted into internal energy residing in the planetary surface materials (*27*). This energy is expended primarily in plastic work related to the finite yield strength of the target rock and in shock heating. Less than 10 percent of the impact energy is converted to kinetic energy of the ejecta escaping from the cavity; even so, a single projectile from a major lunar basin can produce a secondary crater 25 km in diameter (*28*). Thus, in addition to producing the primary crater cavity, the impact involves the redistribution of energy associated with the ejecta to distances several times the cavity radius. Shock waves produced by the impact attenuate into a complex pattern of seismic waves. Although the energy radiated seismically represents less than 1 percent of the total kinetic energy of an event, a large impact may give rise to wave trains with amplitudes several orders of magnitude greater than those from any known terrestrial earthquake, producing major effects on surface materials (*29, 30*). Thus, impact cratering represents an important geological and tectonic process. The process involves vast quantities of kinetic energy (perhaps as much as 10^{34} ergs for the Imbrium basin on the moon) (*30*) delivered to the lunar lithosphere and redistributed in an extremely short period of time (seconds or minutes) to produce major topographical depressions in the lithosphere, uplift and collapse of the crater rim, and significant changes in the physical properties of lithospheric materials.

Craters exceeding about 200 km in diameter characteristically display concentric ring structures in addition to the crater rim, and are termed basins. Basins can exceed 1000 km in diameter and have dramatically influenced the outer parts of the moon. Although the exact depth of excavation and sampling is unknown, there is general agreement that local excavation and mixing during basin formation reached depths of at least several tens of kilometers (*31*). Massive amounts of impact melt are produced in a single basin-forming event, with approximately 200,000 km^3 remaining in the interior of the 900-km-diameter Orientale basin (*32*). Ejecta are emplaced over great radial distances; the edge of the field of recognizable secondary craters from Imbrium lies approximately 90° in arc length (2700 km) from the point of impact (*33*).

The location of the transient cavity and the mechanism by which basin rings are formed are controversial topics. In some models, ring formation is attributed to differential excavation of material as the substrate properties change with depth (*34*), while in others the formation of the outer ring is attributed to collapse of the transient cavity rim just after excavation (*32, 35*). Sublithospheric flow may produce stresses leading to formation of the outer ring by lithospheric failure immediately after cavity excavation (*36*). Also, some basins (Tranquillitatus, Fecunditatis) appear to have undergone extensive viscoelastic relaxation soon after their formation (*37*). The most recent major ringed basin, Orientale, was formed about 3.8 billion years ago, somewhat after the onset of the emplacement of mare basalts. By that time the lithosphere was thick enough to support basin structure. Elevation differences from floor to rim crest are 6 to 8 km (*38*), with the topographically low areas providing locations for the concentration of mare basalt deposits.

Most of the lunar tectonic features are local in scale and are associated with vertical movement in and near mare basins. These features include the linear rilles, generally interpreted as graben (*39*) and most commonly peripheral to major mare basins, and the mare ridges, most of which are within mare units and are plausibly identified as compressive features indicative of crustal shortening (*40*). The spatial distribution of rilles and ridges, the sequence and distribution of major volcanic units, and the topography of the present surfaces of most maria—including those with mascon gravity anomalies—are all evidence for an extended period of mare basin subsidence spanning most of the era of mare volcanism and perhaps continuing until long after volcanism ceased (*41*). The temporal distribution of rille and ridge forma-

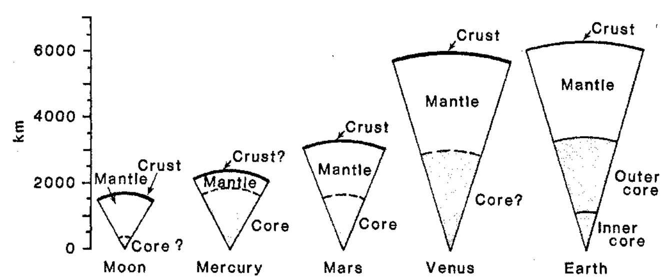

Fig. 3. Interiors of the terrestrial planets. For the moon, a 70-km-thick crust is indicated from seismic refraction, gravity, and topographic data (*16*); only an upper bound on the radius of a possible central core can be obtained from seismic and electromagnetic information (*122*). For Mercury, a core is suggested but not demonstrated from the internal magnetic field and surface geological record (*46, 50, 59*); no information on a crust has been obtained. A distinct crust and core are indicated for Mars from gravity data, but their radial boundaries are poorly defined (*64, 65*). A low-density crust for Venus can be inferred from gravity and lander chemical data (*105, 111*); only the mean density and an assumed similarity to Earth suggest the characteristics of a possible core. The internal structure of Earth is comparatively well known from seismic data (*123*).

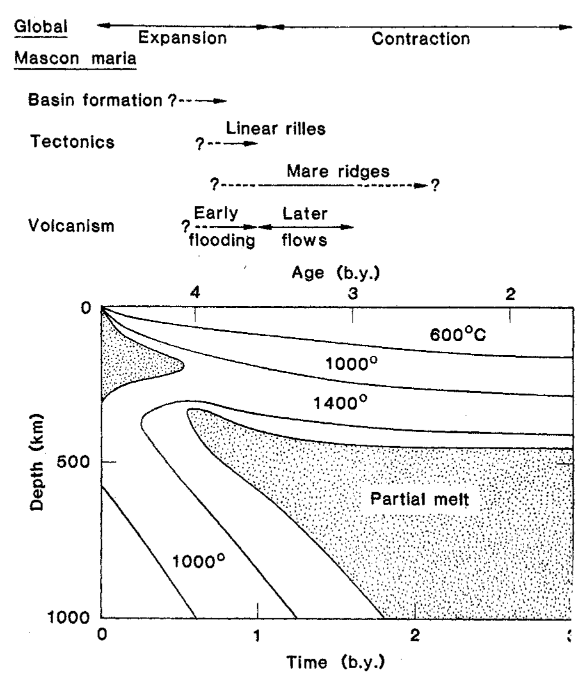

Fig. 4. Summary of temporal relations among lunar thermal history, global expansion and contraction, and the tectonic and volcanic history of mascon maria (*41*). Isotherms are shown versus depth and time for a thermal history model with initial melting to 300 km depth and an initial central temperature of 300°C. Regions of partial melt are shaded; *b.y.*, billion years.

tion, however, is far from uniform. Linear rilles occur only in relatively old mare units and in highlands generally adjacent to maria, and all appear to have formed prior to 3.6 ± 0.2 billion years ago (*42*). Mare ridges, on the other hand, disrupt even the youngest mare basalt deposits, and therefore must have continued to form until less than 3.0 billion years ago.

The spatial and temporal distributions of linear rilles and ridges in the mare regions of the moon appear to be the product of two superposed stress systems: a local stress due to lithospheric loading by basalt fill in the mare basin and a global thermal stress associated with warming and cooling of the lunar interior (*41*). The local horizontal stress radial to the mare basin is compressive beneath the load and extensional at the edge of the load. Under regional extension, the local extensional stress in the outer portions and outside of the mare units is enhanced. Under regional compression, the zone of peripheral extension is suppressed and the central compressive stresses are increased. For a subset of the lunar thermal history models that successfully account for the lack of global-scale extensional or compressive tectonic features, an early period of modest global expansion and lithospheric extension is confined to the first 1.0 billion years of lunar history. This extensional period is followed by a more extended period of modest planetary contraction and lithospheric compression lasting until the present (Fig. 4). These models account for cessation of linear rille formation by the onset and gradual increase of global horizontal compressive stress associated with the period of lunar contraction.

With plate flexure theory, the geometry of basalt fill and sequence of mare filling can be combined with information about the timing and location of tectonic features in order to derive estimates for the thickness of the elastic lithosphere of the moon at the time those tectonic features were formed (Fig. 5) (*41, 43*). There is strong evidence for an increase in the effective thickness of the elastic lithosphere between ~ 0.1 billion years after basin formation and the time when volcanic filling of maria ceased. Lithospheric thickness was sufficient subsequent to mare volcanism to have maintained much of the final load as evidenced by the mascon gravity anomalies. The thickness of the lithosphere 3.6 to 3.8 billion years ago was markedly heterogeneous, as indicated by the varied tectonic response to imposed loads. Thickness at that time varied from more than 75 km for the eastern basins Nectaris, Crisium, and Smythii to as little as 25 km near Grimaldi and western Procellarum. These variations in lithospheric thickness are a direct reflection of variations in the shallow temperature structure of the moon. Regions of thin lithosphere are correlated with areas of high surface concentrations of radioactive elements, of late-stage nonmare volcanism, and of the youngest centers of mare volcanic eruptions (*44*).

In summary, the moon shows little or no direct evidence in its surface features for global stresses sufficient to have caused widespread lithospheric failure. At least since the formation of a thick global crust, the moon has been a one-plate planet. Its early history was dominated by impact bombardment, mare volcanism, and global lithospheric stress associated with mild expansion followed by mild contraction of the interior. The observed tectonic features are the result of stresses produced by local loading by mare basalts superposed on the global stresses related to interior heating and cooling. The local loads caused vertical tectonic movement (downwarping) of the single lithospheric shell. Variations in lithospheric thickness in time and space influenced the style and extent of vertical tectonics.

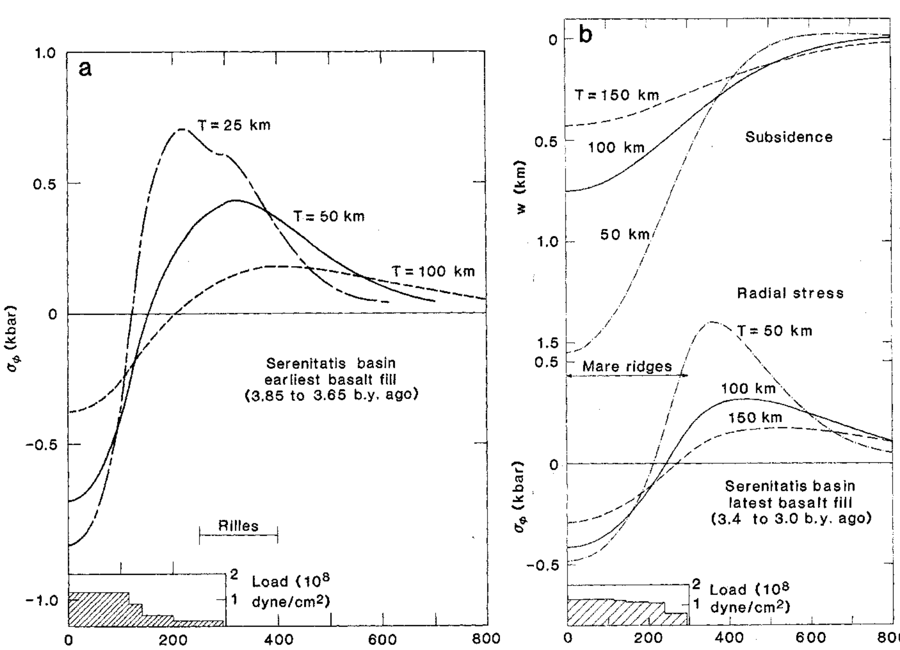

Fig. 5. Evolution of lithospheric thickness in the Serenitatis basin region (*41*). (a) Surface horizontal stress component σ_ϕ (extension is positive), radial with respect to the basin center, for a model of the superisostatic load (inset) in Serenitatis at a time just after earliest basalt fill. The radial extent of concentric rilles is indicated by the bar. (b) Subsidence w and radial stress σ_ϕ for load model (inset) in Serenitatis at a time just after emplacement of the youngest basalt fill. The radial extent of mare ridges is indicated by arrows. Curves are shown in (a) and (b) for several possible values of elastic lithosphere thickness T. A value of 45 ± 15 km for T provides a best fit in (a); 125 ± 25 km provides a best fit in (b); *b.y.*, billion years.

Mercury

The surface of Mercury, like that of the moon, has been heavily modified by impact craters and basins (Fig. 1). Of the 40 percent of the planet viewed by Mariner 10, about one-fifth of the surface is composed of smooth plains comparable to the lunar maria both in gross morphology and in their younger age compared to adjacent terrain (*45*). Despite these similarities, differences exist. Relatively large areas of old intercrater plains are seen on Mercury, and all plains units have an albedo similar to that of the older cratered highlands. There is little evidence in either the smooth or the intercrater plains for structures comparable to volcanic constructs, flow fronts, or vents as seen on other planets. Thus, the hypothesis of a volcanic origin of mercurian plains units is controversial (*46, 47*). In addition, large lobate scarps up to several hundred kilometers long (Fig. 6) have modified all terrain units on the mercurian surface (*46, 48*). Finally, the density of Mercury (5.4 grams per cubic centimeter) implies the presence of about 60 to 70 percent iron (by mass) in the interior (*49*). The high iron content and the presence of an internally generated magnetic field (*50*) are consistent with the hypothesis that Mercury has a large metal core that is at least partly fluid (Fig. 3). Thus, while the surface of

Mercury is similar to that of the moon, the deep interior may be more comparable to that of Earth.

The areal density and size distribution of impact craters in the heavily cratered terrain of Mercury have many similarities to those of craters in the lunar highlands. These similarities have led to the hypothesis (51) that at one time both bodies (and, by inference, the other terrestrial planets) were subjected to a high flux of impacting objects and that this high flux ended at the time of formation of the youngest large lunar basins about 3.8 billion years ago (52). The hypothesis is supported by dynamical studies of the possible populations of objects that could have been responsible for such a late heavy bombardment of the inner solar system (53). The end of heavy bombardment, by this hypothesis, provides a "marker horizon" for the geological history of the terrestrial planets: heavily cratered surfaces are older than 3.8 billion years, while lightly cratered surfaces are younger.

The lobate scarps on Mercury have been interpreted as thrust or reverse faults on the basis of morphology, transection relations, and the foreshortening of at least one crater cut by a younger scarp (Fig. 6) (46, 48). Formation of the scarps extended over a time interval from near the end of heavy bombardment until after the emplacement of the smooth plains. Because no instance of embayment of scarps by intercrater plains is observed, scarp formation may have begun after the emplacement of the intercrater plains (54).

Two processes have been proposed to account for the lobate scarps: tidal deceleration and global contraction (46, 55). Mercury's tidally evolved spin-orbit resonance (56) shows that despinning has occurred, but whether the present resonance was achieved before or after the period of scarp formation is unknown. The tidal despinning hypothesis for scarp formation leads to the predictions of a preferred orientation for near-equatorial scarps and of complementary extensional features in the vicinity of the poles (55). Whether there is a preferred orientation for lobate scarps is the subject of debate (57), but no evidence has been found for near-polar extensional tectonics. Thus, even if despinning were an important source of stress during scarp formation, a superposed global compressive stress due to planetary contraction would be required to account for the observed tectonic patterns.

The amount of global contraction necessary to account for the geometry and distribution of scarps is 1 to 2 km of planetary radius (46, 58). On the basis of thermal history models for Mercury which include the effects of core differentiation and subsequent planetary cooling, the timing and extent of planetary contraction place important limits on Mercury's interior evolution. Core formation must predate the end of heavy bombardment near 3.8 billion years ago, because core-mantle differentiation is accompanied by the release of a large amount of gravitational potential energy as heat and by a substantial (15- to 20-km radius) expansion of the planet (59). The effects of such heating and expansion are not evident in the preserved tectonic features on Mercury, although the extensive volcanic resurfacing that may have accompanied core infall may still be recorded in the ancient intercrater plains. Subsequent to planetary differentiation, cooling of the planet, perhaps augmented by partial solidification of a fluid core, could account for global contraction of the magnitude observed (60). The contraction recorded by the lobate scarps would be accompanied by horizontal compressive stress in the lithosphere. The extensional stress conducive to eruption of the smooth plains (if volcanic) may therefore have been local in origin, probably associated with impact basin formation and topography. The increasingly compressive global stress would have tended to shut off volcanism by overriding local extensional stress. Thus the duration of smooth plains volcanism on Mercury is predicted to be shorter than the duration of mare volcanism on the moon (8).

The largest impact structure observed by Mariner 10 is the Caloris basin (1300 km in diameter), which is filled with smooth plains of impact melt or volcanic origin. The floor of the basin is laced with an extensive pattern of concentric and radial fractures and ridges. The fractures become wider and deeper toward the basin center and generally cut the ridges. These relations have led to the proposal that the ridges are compressional features formed by basin subsidence and that the fractures are extensional features formed by subsequent uplift of the basin floor (46). Initial subsidence has been attributed to magma withdrawal, and subsequent uplift to isostatic readjustment. Caloris is located near the axis of the minimum moment of inertia, and it has been suggested that there is an annular positive gravity anomaly associated with uncompensated smooth plains surrounding the basin (61). This positive anomaly, rather than a mascon associated with the interior of the basin, may account for the coincidence of Caloris and the long axis of the dynamical figure of the planet. The re-

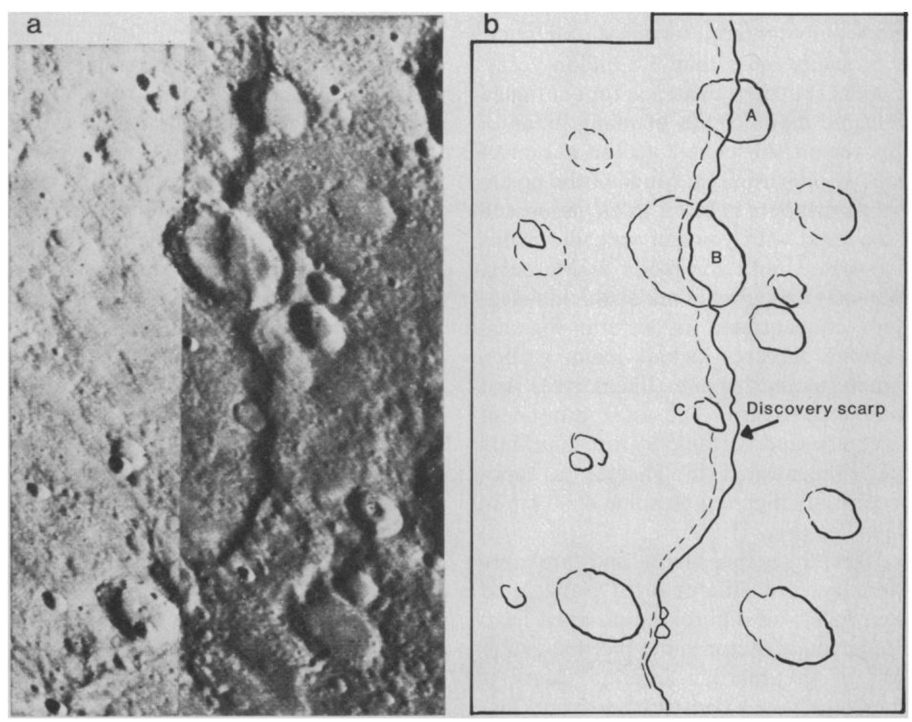

Fig. 6. Lobate scarp on Mercury. (a) Mariner 10 images (FDS 27398, 27399); (b) sketch map of same area. Discovery scarp is one of the most prominent lobate scarps observed by Mariner 10. It is interpreted as a thrust or reverse fault, indicative of lithospheric compression. The scarp is about 550 km in length and cuts two craters (*A*, 35 km in diameter, and *B*, 55 km in diameter). The maximum height of the scarp, about 3 km, occurs in the area south of crater B. The sense of faulting is an overthrust of the left side over the right (46, 54).

sponse of Mercury's lithosphere to such an annular load includes uplift and extension within the Caloris basin, thus providing an explanation for the fractures on the basin floor (*62*).

In summary, the early history of Mercury was dominated by a high rate of bombardment accompanied by basin formation. An initial phase of global expansion associated with planetary differentiation may have favored volcanism coincident with the late heavy bombardment. Basin filling and associated tectonics are distinct from the style observed on the moon. Cooling and contraction caused global compressional stresses sufficient for disruption (but not destruction) of the single lithospheric plate and formation of the extensive system of scarps. Scarp distribution patterns may have been affected by tidal despinning of the planet. The extensive global compression was likely to have restricted surface volcanism to a period shorter than that for the moon.

Mars

The Mariner 9 and Viking missions to Mars revealed a planetary surface more geologically complex than that of the moon or Mercury (Fig. 1) (*63*). Although a roughly hemispherical region of the martian surface is composed dominantly of cratered terrain and basins and, like the heavily cratered terrain on Mercury, is probably older than 3.8 billion years, there is abundant evidence for continued volcanic modification of many areas of this region subsequent to the period of heavy bombardment. Much of the northern hemisphere is lower in elevation and is covered with younger volcanic plains deposits. Unlike the moon or Mercury, Mars has a number of major shield volcanoes concentrated in several regions. Tectonic features include major graben structures and smaller linear rilles and mare-type ridges. The most prominent structural and volcanic features on Mars are concentrated in Tharsis, a topographically high region some 4000 km in diameter (Fig. 7).

Mars, like the moon and probably Mercury, is a differentiated planet, and the process of differentiation must have been a major factor in the thermal evolution of the martian interior. Mars is known to have a dense central core (Fig. 3) by reason of its moment of inertia (*64*). Mars also has a distinct low-density crust of variable thickness or density, as indicated by the partial to complete isostatic compensation of surface topography (*65*). Since isostasy prevails in the ancient, heavily cratered uplands, the martian crust must have been formed prior to 3.8 billion years ago. By analogy with the moon, an early hot exterior for Mars is a likely prerequisite for such early crustal differentiation.

The tectonic history of Mars differs from that of Mercury and the moon. In contrast to Mercury, there are no large-scale thrust or reverse faults indicative of global contraction. On the contrary, all of the large-scale tectonic features are extensional. As on the moon and Mercury, there are small tectonic features, both extensional and compressional, that are the product of some combination of local and global stress fields. Thus, any contraction phase in the evolution of Mars must have involved at most a minor amount of horizontal compressive stress, and global expansion and lithospheric extension is favored over most of the planet's history. An extended period of plains and shield volcanism over much of the martian surface, lasting perhaps until geologically recent times, also supports a planetary history dominated by lithospheric extension (*8*).

The thermal history models that most readily match these constraints involve, as for the moon, a hot exterior and a cold deep interior as initial conditions. Compared to the moon and Mercury, a greater fraction of the volume of Mars is postulated to have been at low temperatures initially, allowing interior warming and global expansion to continue until geologically recent times (*66*). Thus, while crustal formation on Mars, as on the moon, would have been an early consequence of the initial heating and melting of the exterior, differentiation of the deep interior and completion of core infall may have occurred later than in the other terrestrial planets.

Following crustal formation, a global lithosphere developed that, as on the moon, thickened with time. Both crust and lithosphere were sufficiently thick by the end of heavy bombardment that Mars has since been a one-plate planet lacking the potential for lithosphere subduction and therefore for plate tectonics. The youngest large basins on Mars have free-air gravity anomalies, suggesting incomplete isostatic compensation of basin topography and therefore a lithosphere of finite strength (*67*). Distinct positive gravity anomalies are seen over the Isidis "mascon" basin and over a number of the large volcanic shields (*68*). Persistence of these gravity anomalies for geologically long times also requires a martian lithosphere of substantial strength. Evidence for a generally thickening martian lithosphere includes the observation that the heights of volcanic constructs on Mars show a negative correlation with surface age, suggesting an increase in magma hydrostatic head with time due to a deepening of magma source regions (*69*).

The thickness of the elastic lithosphere beneath major volcanic loads may be estimated on Mars, as on the moon, by the tectonic response to loading—in particular, by the radial distance of circumferential graben around lava-filled basins and large volcanic constructs. Based on the observed positions of graben in the Isidis basin area and surrounding the major martian shields, a heterogeneous lithosphere is indicated, ranging in thickness from 25 km to greater than 150 km (*70*). The values show no correspondence with the surface age of the load features; loads with both young (Olympus Mons) and old (Isidis plains fill) surfaces give large values of elastic lithosphere thickness. The thinnest elastic lithosphere is indicated for the central portions of the major volcanic provinces Tharsis and Elysium, suggesting that locally thin lithosphere and anomalously prolific volcanism are closely related.

Tharsis is the most elevated region of the planet (Fig. 7). The broad topographic high stands about 10 km above surrounding terrain, and heavily cratered units are exposed at great elevations. Swarms of graben surround Tharsis in a crudely radial array (*71*); Valles Marineris is a large graben system extending radially away from central Tharsis. The crest and flanks of Tharsis are dotted by several major shield volcanoes often rising some 25 km above local elevation. The traditional explanation of Tharsis is that broad updoming of the lithosphere caused by a thermal or chemical anomaly in the mantle led to fracturing and to the volcanic emplacement of thin plains units and later of the large shields (*71, 72*). Evidence cited in support of this model includes the broad topographic high of Tharsis, the large elevation of surface units mapped as relatively old on the basis of crater and fracture density, and the generally radial trends of most extensional fractures in and near Tharsis.

An alternative explanation for the events of Tharsis was recently proposed (*73*). In briefest terms, the explanation contains three elements.

1) The elastic lithosphere of Mars has been heterogeneous in thickness (that is, in temperature) throughout much of its history. Stress due to both global and

local causes was concentrated in zones of thin lithosphere, foremost among which was (perhaps a small fraction of) Tharsis. Thus, fracturing was concentrated in such areas.

2) Intense fracturing was conducive to local volcanism. The heating associated with volcanism maintained a locally thin elastic lithosphere, ensuring that further fracturing and volcanism would be concentrated in the same area.

3) The topography of Tharsis was produced primarily by volcanic construction. The topographic and gravity highs have been supported against complete isostatic compensation by the strength of a broadly thick (if locally thin) lithosphere.

A major distinction between this explanation for Tharsis and earlier explanations is that no anomalous properties need to be attached to the mantle beneath major martian volcanic provinces. The location of volcanism on Mars is governed primarily by the sites of easiest access of magma to the surface. In this respect, Tharsis may be similar to mid-ocean ridges on Earth. The Mid-Atlantic Ridge and the East Pacific Rise are major volcanic centers and are nearly stationary in a hot spot reference frame (*74*), yet no anomalous characteristic is attributed to the mantle beneath them. The mantle beneath ocean ridges, and beneath Tharsis, plays a passive rather than an active role in this scenario.

In summary, an extended period of volcanism for Mars is consistent with the lack of major global compressional features on the surface. Lateral variations in a generally thick lithosphere caused local concentration of volcanic activity on the single lithospheric plate. Tectonism was vertical rather than lateral. At least some of the major basins were loaded and deformed in a manner analogous to those on the moon. The Tharsis province may be related to anomalous mantle regions and consequent massive vertical uplift and volcanism or, conversely, to a concentration of faulting in a locally thin lithosphere and associated large-scale volcanic construction.

Earth

Earth stands apart from the other terrestrial planets in several respects, notably in the presence of vast oceans of liquid water and in the occurrence of conditions of surface temperature and atmospheric composition conducive to the development and evolution of life. Earth also differs profoundly from the smaller terrestrial planets in its plate tectonics, which involves the large-scale recycling of lithosphere into the underlying mantle, a dichotomy between continental and oceanic lithosphere, and the formation of great mountain belts where oceans close and continents collide (*75*). The continuous motions of Earth's plates account for most of the world's earthquakes, and the boundaries between separating or converging plates are the sites of most of its volcanic activity (Fig. 8).

Plate tectonics owes its principal characteristics to the lateral rigidity of terrestrial lithospheric plates. Because of that rigidity, plates with horizontal dimensions of thousands of kilometers move as distinct mechanical units. The finite strength of the lithosphere is an important aspect of the subduction process, wherein wholesale sinking of the lithosphere at deep-sea trenches recycles oceanic crust and upper mantle into the deep interior. The subduction process, driven by the negative buoyancy of the slab of sinking lithosphere, may be the dominant driving mechanism in the plate tectonic cycle (*76*).

On Earth, as on other planets, there are large and systematic variations in the thickness and rheological properties of the lithosphere, variations governed principally by the thermal structure of the oceanic mantle. The effective elastic lithosphere in oceanic regions thickens progressively with age of the sea floor (Fig. 9), a consequence of the steady cooling of the spreading oceanic plate (*77*). The deepening of the sea floor with age is the most direct measure of this cooling of oceanic lithosphere (*78*). Continental lithosphere is somewhat thicker than that in the ocean basins (*79*), probably due to generally lower temperatures in the top few hundred kilometers (*80*). The lithospheric thickness in continents is also dependent on time and location, with areas of recent tectonic activity having higher temperatures and thinner

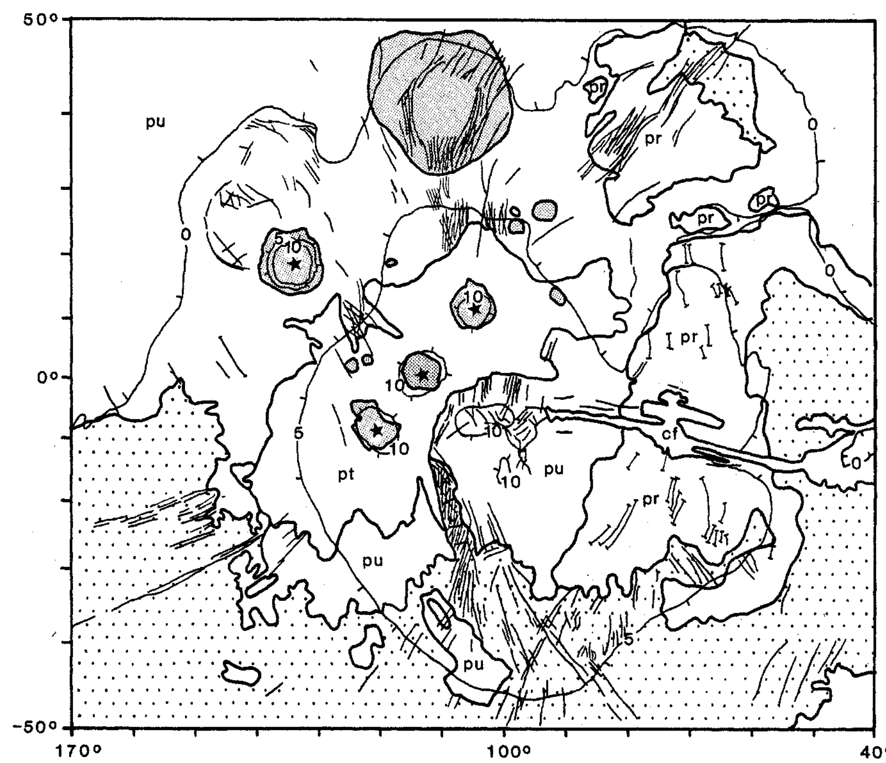

Fig. 7. The Tharsis region of Mars. The region forms a bulge approximately 4000 km in diameter. The topography, indicated by contour lines at 5-km intervals (ticks point downslope), rises more than 10 km above Mars datum. Ancient cratered terrain, indicated by the lighter shading, surrounds Tharsis to the south and is also exposed at high elevations within Tharsis. Plains units are primarily of volcanic origin. Ridged plains (*pr*) are characterized by the abundance of mare-ridge structures. Undivided plains (*pu*) make up the bulk of surface units and extend to the north of Tharsis. Younger volcanic plains (*pt*) surround the major central shield volcanoes. Major volcanic edifices and structures are indicated by the darker shading. Four of these, indicated by stars at their summits, exceed 25 km in elevation. Tectonic features are abundant, with the major Valles Marineris system extending several thousand kilometers from central Tharsis toward the east (*cf*, canyon floor material). Linear rilles (lines) are arrayed in a crudely radial pattern, while mare ridges (lines with ticks) are arrayed in a more concentric pattern around Tharsis. Width of diagram is about 4000 km at the equator. Data for map are derived from (*72*) and (*124*).

lithosphere than the stable continental shields (*81*).

The local thickness of the lithosphere plays an important role in the processes leading to vertical tectonics, including loading by volcanic or sedimentary deposits and rifting and subsequent subsidence. Volcanic loads in ocean basins produce a flexure of the oceanic plate, the magnitude and dominant wavelengths of which are controlled primarily by the effective thickness of the elastic lithosphere. Loads on young sea floor are locally compensated except at wavelengths shorter than about 100 km (*82*), while loads on older lithosphere, such as the Hawaiian island chain (Fig. 10), produce a broad depression flanked by an outer bulge, indicative of an elastic plate some tens of kilometers thick (*83*). The deformation of Earth's lithosphere by volcanic loads is a direct analog to the loading of a planetary lithosphere by basalt fill in a large basin, as on the moon, or by a volcanic construct, as on Mars.

The signature of the lithosphere is also evident in the rifting apart of two continental masses and the subsidence of continental margins and interior continental basins. Opening of a new ocean basin appears to be initiated with the stretching and thinning of continental lithosphere (*84*). Subsidence of the rifted lithosphere is caused by a combination of thermal contraction and loading by igneous intrusions and later sediments (*84, 85*); regional adjustment by lithospheric flexure is an important aspect of the adjustment to those loads (*86*). Large sedimentary basins within continents are in many ways analogous to continental margins, since both thermal contraction and lithospheric flexure in response to sedimentary loading appear to have occurred (*87*).

The geological record preserved at Earth's surface is dominantly a product of its plate tectonics. The oceanic crust is everywhere younger than about 200 million years, testimony to the efficiency with which oceanic lithosphere is recycled into the deeper mantle. The continental crust is more heterogeneous in both composition and age. The oldest preserved crustal rocks date from about 3.8 billion years ago (*88*), and rocks from the stable interiors of most continents are found with ages of at least 3.5 billion years (*89*).

Recognition of the roles of plate subduction and continental collision in the creation of presently active mountain ranges has led to the realization that many older mountain belts were also produced by plate convergence and the closing of an ocean basin by lithosphere subduction (*90*). Much of the history of the continental crust can be read as the consequence of a series of "Wilson cycles," each marked by a period of continental rifting and the opening of an ocean by sea-floor spreading followed by a period of subduction, ocean closure, continental collision, and mountain building (*91*).

Differentiation of Earth (Fig. 3) has progressed on a variety of spatial and temporal scales. Perhaps most rapid was the separation of the core from the mantle. Both U/Pb isotope characteristics (*92*) and paleomagnetism of ancient rocks presumably magnetized by an internal field produced in a fluid core (*93*) reveal that the bulk of core-mantle differentiation occurred in the first few hundred million years of Earth's history, and core infall during the later stages of accretion is a distinct possibility (*94*). Crustal differentiation took longer. The 3.8-billion year apparent upper limit on radiometric ages of crustal rocks (*89*) and the indication from Sm-Nd isotope systematics that crustal rocks older than 2.5 billion years may have been derived from unfractionated mantle reservoirs with chondritic relative abundances of rare earth elements (*95*) suggest that little or no stable continental crust appeared on Earth until several hundred million years after planetary formation. The volume of continental crust has probably grown more or less continuously with time; radiometric age data indicate a possible peak in the production rate near 3.0 billion years ago (*89*). On yet a third time scale, isotopically distinct source regions of oceanic and continental vol-

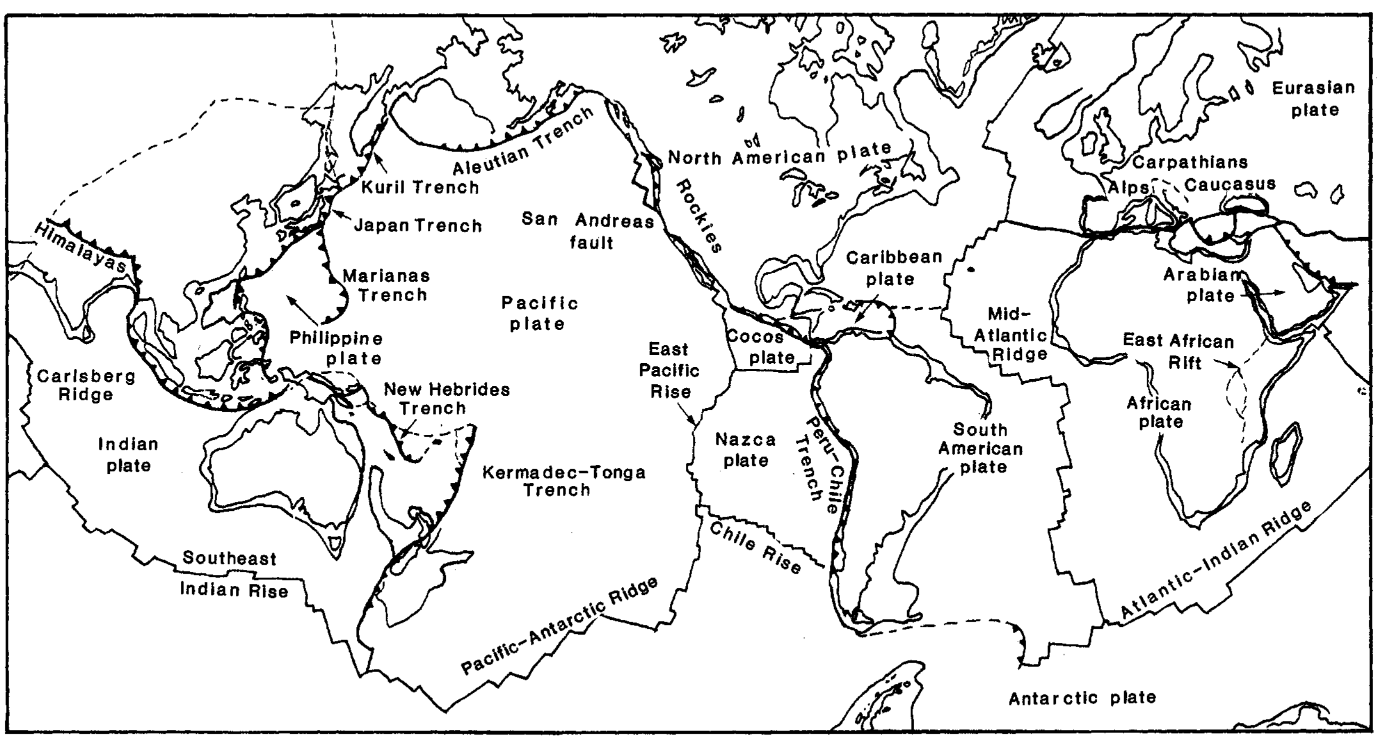

Fig. 8. Tectonic map of Earth showing boundaries of major lithospheric plates (Mercator projection) (*125*). Coastlines and continental shelves are outlined. Plate boundaries are indicated by solid lines where certain and by dashed lines where uncertain. Trenches are shown by hachured lines, with hachures on the overriding plate.

canic rocks have persisted for billions of years in Earth's mantle despite the mixing effects of mantle convection (*96*). Whether these distinct magma reservoirs indicate chemical and isotopic layering of the mantle (*97*) or lateral heterogeneities on a smaller spatial scale (*98*) has not been resolved.

Thermal history models for Earth generally are based on the premise of initially high temperatures. The grounds for this premise include the evidence for early high temperatures in the other terrestrial planets, the large energy input expected from the late stages of accretion (*94*), and the substantial additional source of heat from core infall (*99*). A consequence of such a large budget of initial heat would be a vigorously convecting mantle, a large surface heat flux, and a generally thin and probably easily deformable lithosphere early in Earth's history. There is some indication that this early heating is still supplying a significant fraction of the heat flow (*100*).

Both the volume of permanent crust and the mean thickness of the lithosphere probably increased slowly with time. As noted above, radioisotope systematics and measured ages of continental rocks both suggest that a permanent crust may not have formed until several hundred million years after planetary formation (*89, 95*). Vigorous convection and impact stirring may have served to quickly remix any protocrust back into the underlying mantle. The 3.8-billion-year age of the oldest known terrestrial rocks is sufficiently close to the time at which the heavy bombardment of the inner solar system ceased (*52*) to suggest that impacts may have played a role in delaying stabilization of a terrestrial crust (*101*).

The thickening of the lithosphere was a necessary prerequisite for the emergence of plate tectonics in its present form. The subduction of oceanic lithosphere and the laterally coherent motions of large plates are both tied to the finite elastic strength of an outer layer several tens of kilometers thick. When the lithosphere first became sufficiently thick on Earth is uncertain, but both the geological record (*91*) and estimates of ancient continental geotherms (*102*) are consistent with the view that plate tectonics in approximately its present form has been an essential aspect of the thermal evolution of Earth for the past several billion years.

In summary, the evolution of Earth, in distinct contrast with that of the smaller terrestrial planets, has been marked by large-scale lateral motions and wholesale recycling of lithospheric plates. The plate tectonic cycle has dominated the heat budget of the outer several hundred kilometers of Earth and has led to a level of volcanic activity which, when integrated through time, surpasses by a huge factor that seen on the moon, Mercury, or Mars. Earth shares with the other terrestrial planets, however, the concept of the lithosphere as an important governor of tectonic style. The vertical tectonics on Earth associated with volcanic loading, continental rifting, and subsidence of continental margins and platform basins has many direct analogs on the other planets.

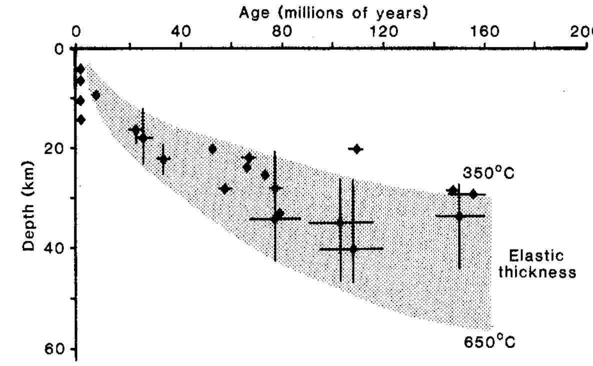

Fig. 9. Estimates of the elastic thickness of Earth's oceanic lithosphere as a function of age, based on the flexural response to applied loads (*126*). The shaded area represents the region of oceanic lithosphere with temperatures between 350° and 650°C (*78*).

Venus

Venus, like Earth, is approximately twice the diameter of Mars, and thus might be expected to be more comparable to Earth than to the smaller terrestrial planets. Venus also shows strong similarities to Earth in terms of density and distance from the sun. However, Venus remains the most poorly characterized terrestrial planet. Its dense, cloud-filled atmosphere has precluded a view of the surface features at a resolution sufficient to characterize its geological nature and evolution.

The Pioneer mission to Venus (*103–105*), in conjunction with Earth-based observations (*106*), provided at low resolution a picture of the large-scale topography and gravity field of the planet. The Pioneer Venus radar altimeter has mapped the topography of much of the surface of Venus at about 100-km horizontal resolution (approximately equivalent to 1° squares on Earth and Venus) (Fig. 11). The global topography of Venus shows a unimodal elevation-frequency distribution, in contrast to the bimodal distribution of Earth caused by the distinct topographic levels of continents and ocean basins (*103*). Approximately 60 percent of the mapped surface lies within 500 meters of the modal altitude. The terrain of Venus (Fig. 11) has been subdivided into lowlands (20 percent), rolling plains (70 percent), and highlands (10 percent) (*104*).

The highlands are unlike any topographic feature seen on the smaller terrestrial planets. Ishtar Terra, larger than the continental United States, stands several kilometers above mean planetary radius and is separated from the surrounding rolling uplands by relatively steep topographic gradients; the western part is a vast plateau 2500 km in diameter (Lakshmi Planum) standing 3.5 to 4 km above mean radius. Mountain ranges occur irregularly along the plateau edges, rising as high as 3 km above the central plateau plains. Eastern Ishtar is more irregular in outline and contains Maxwell Montes, the single most dramatic topographic feature on Venus. These mountains are centrally located and rise up to 11 km above the mean radius, a height exceeding that of Mount Everest above Earth's sea level. Lakshmi Planum is at an elevation above mean radius comparable to that of the Tibetan Plateau above sea level but covers approximately twice the area. Aphrodite Terra, about the size of Africa, is the largest highland region on Venus and extends along the equator for over 10,000 km. Aphrodite appears topographically rougher and more complex than western Ishtar (Lakshmi Planum) and contains a large, roughly circular structure about 2400 km in diameter. A number of dramatic linear depressions up to 3 km deep and several hundred kilometers wide transect central eastern Aphrodite; some are more than 1000 km long and possess raised rims.

The lowland areas of Venus are located in roughly circular areas, such as the large depression east of Ishtar, and in very broad but linear depressions, such as that extending along the southern edge of Ishtar. These regions are relatively smooth and make up less than 20 percent of the planet's surface. The extensive midland regions, with elevations near the planetary mean, contain a diversity of topographic features revealed by both Pioneer Venus and Earth-based radar data, including linear troughs, parallel ridges and troughs, and numerous shallow circular structures often containing central mounds. These latter features, ranging in diameter from 20 to 1700 km, have been interpreted as impact craters (*107, 108*) and as the surface manifestations of igneous intrusions (*109*). Understanding of the origin of these features is critical. If they are impact craters, their presence implies an age of several billion years for portions of the surface of Venus.

Our knowledge of the terrain of Venus thus strongly hints at a diversity of geological processes, including tectonism, volcanism, and impact cratering (*107, 110*). Although the surface of Venus has become much better known in terms of large-scale physiography, the resolution is still insufficient to characterize the geological processes responsible for the diverse and tantalizing topography.

Little is known about the interior structure and composition of Venus or its thermal evolution (Fig. 3). Evidence for a chemically distinct, low-density crust includes at least partial isostatic compensation of highland topography (*105*) and the results from Venera landers (*111*) indicating low bulk density and high concentrations of radioactive elements in surface rocks. Although Venus has a mean density similar to that of Earth, the existence of a central metallic core on Venus is unproven. In striking contrast to Earth, Venus has no measurable magnetic field of internal origin (*112*).

Pioneer Venus measurements of the chemical and isotopic abundances of rare gases in the atmosphere have provided important information on the volatile inventory and evolution of the planet (*113*). The abundance of atmospheric ^{40}Ar is about one-third that of Earth. Argon-40 is the product of radioactive decay of ^{40}K and reaches the atmosphere from the interior of a planet by volcanic and tectonic processes and surface weathering. Although this observation could be explained by different initial abundances or by different rates of liberation from surface rocks due to different weathering processes, it may also hint that, assuming similar initial abundances, volcanic and tectonic processes have not been as efficient in producing new material at the surface of Venus as on Earth. The abundance of nonradiogenic rare gases in Venus's atmosphere, on the other hand, exceeds that of Earth's atmosphere by a factor of about 100 (*113*). In concert with the very low abundance of these gases in the martian atmosphere (*114*), an explanation for the

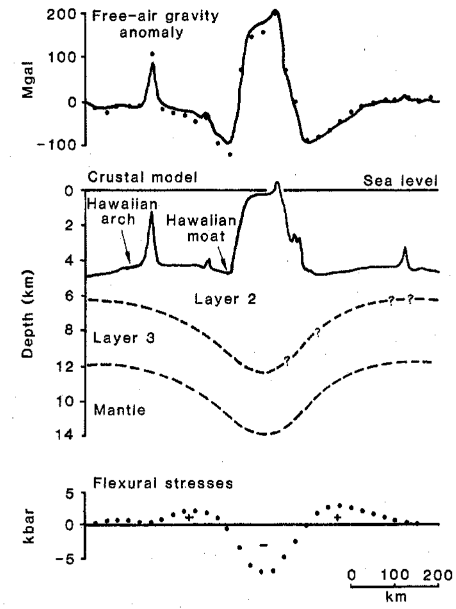

Fig. 10. Free-air gravity, crustal model, and flexural stress for the Hawaiian island chain (*126*). The flexural model (center) gives the surface bending stresses (bottom) and the predicted free air-gravity (dots) compared with observations (solid line) at top.

volatile inventory of the terrestrial planets involving only the condensation of planet-forming materials in a primitive solar nebula with strong radial temperature gradients (*4*) is clearly insufficient. Modifications to models of the early thermal structure of the nebula (*114*) and either a cometary or solar origin for the nonradiogenic rare gases (*115*) have been proposed as alternative explanations.

Another important characteristic of Venus which bears on its internal and surface evolution, relative to that of the Earth, is its higher surface temperature (450°C greater than Earth's). Such a temperature should result in more easily deformable near-surface materials (*116*) and perhaps in a greater degree of partial melting and more efficient separation into the crust of the basaltic component of the mantle (*117*). Limited water in near-surface rocks may also be significant in providing strength and rigidity to the outer parts of the planet (*118*). The influence of higher surface temperatures on the near-surface rheology may be balanced by the virtual absence of water, producing an Earth-like lithospheric thickness (*116, 118*). The net result of several of these factors could be a thicker crust than in terrestrial ocean basins and therefore a lower density lithosphere that could inhibit lithospheric subduction and recycling (*117, 119*).

Tracking of the Pioneer Venus spacecraft near its closest approach to the planet has provided information on the gravity field of Venus and has shown that the spectra of the gravity field and of surface topography are positively correlated (*105, 120*). Gravity anomalies are comparable in magnitude to those observed on Earth and are smaller than those on the moon and Mars. The high level of correlation between long-wavelength gravity anomalies and topography strongly argues for a relation between their processes of origin. Some possible explanations (*121*) include: (i) Venus topography is compensated but at great depth; (ii) the lithosphere is thick and topography is only partially compensated, with the remainder of the support derived from lithospheric strength; and (iii) the lithosphere is thin and the topographic variations are dynamically supported by stresses associated with mantle convection.

The observed gravity anomalies, although correlated with topography, are much smaller than the anomalies predicted from the topography alone, implying that a significant fraction of the topographic variations are isostatically compensated (*105, 120*). The compensation depth appears to be less than about 100 km (*105, 121*), although it is not known whether the density difference providing the compensation is compositional or thermal. Possible explanations for the correlation between long-wavelength gravity and topography cannot be distinguished on the basis of available data, so the thickness of the lithosphere remains uncertain.

In summary, the surface of Venus appears to have crater-like features similar to those on the ancient lithospheres of small, one-plate planets but also has rifts and plateaus analogous to structures on Earth. Several models of the interior and thermal evolution of Venus have been proposed to account for various interpretations of the physiographic features. Some emphasize an Earth-like nature and evolution (*109*), others stress similarities to the smaller, one-plate planets (*104*). These models notwithstanding, we do not possess images of sufficiently high image or topographic resolution to accurately characterize geological processes and the evolution of surface topography on Venus. Thus we are faced with many questions. What is the origin of the highland plateaus, such as Ishtar Terra? Are they comparable in origin and composition to continents on Earth, or do they represent a dynamically supported uplift? Are the major mountains, such as Maxwell Montes, of plate tectonic origin, or do they represent massive localized outpourings of lava comparable to intraplate volcanism on Earth or to the

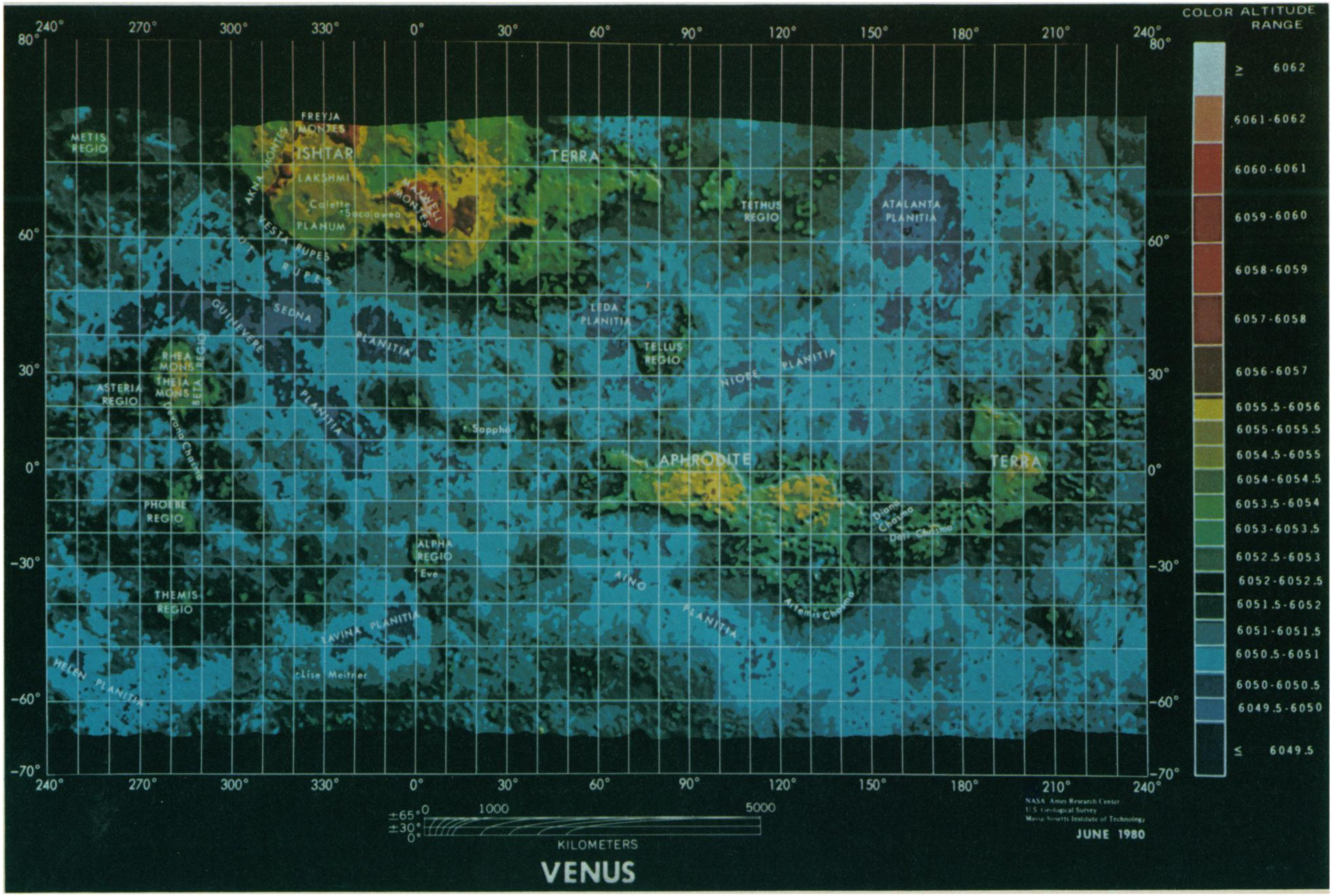

Fig. 11. Topography of the surface of Venus (Mercator projection). The large highland area in the north, Ishtar Terra, is capped by Maxwell Montes (60°N, 0°), which rises more than 13 km above lowest elevations on Venus. Aphrodite Terra straddles the equator from 60° to 210°E (*103, 104*).

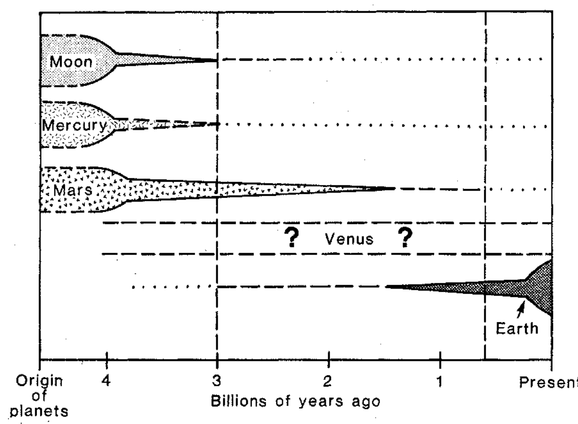

Fig. 12. Ages of planetary surface units. The area plotted for each planet is an estimate of the percentage of the present surface area of different ages (Fig. 1). For the moon, 80 percent of the present surface was formed in the first 600 million years of lunar history, while less than 20 percent (mare basalts) was emplaced between 3 and 4 billion years ago. Since that time, only minor deposits (primarily impact craters) have been formed (represented by dots). The record for Mercury appears comparable to that for the moon (*54*). The cratering flux is uncertain at Mars (*63*), and this results in uncertainties in the percentage of surface area formed between 1 and 3 billion years ago. Although volcanism appears to have extended well beyond the end of similar activity on the moon, the vast majority of exposed surface units were formed in the first half of solar system history. On Earth, over two-thirds of the surface (ocean basins) was formed less than 200 million years ago and surface rocks older than 3.5 billion years are rare. The ages of surface units on Venus are unknown. Crater-like features suggest that some regions may be ancient, while plateaus such as Ishtar and the Maxwell Montes hint at youthful structure.

great shield volcanoes on Mars? Did the circular depressions on Venus originate from meteoroid impact? How did the crust of Venus evolve? Answers to these questions must await the further exploration of Venus.

Discussion

A basic aspect of planetary history is the mode by which a planet generates and then rids itself of heat. Sources of internal and external energy which provide heat include energy associated with accretion and subsequent bombardment, differentiation, tidal interaction, and radioactivity. The budget of these sources over the history of the planet helps to define its thermal evolution. The thickness and other properties of the lithosphere of a planet are closely controlled by both the planetary heat flow and the mechanism by which the planet loses this heat. The geological history of the surface provides evidence as to the nature of the lithosphere through geological time and is thus a key to understanding the thermal evolution of a planet. For example, massive extrusive volcanic deposits bear testimony to rifting of the lithosphere and the direct transfer of heat to the surface. As a second example, a change over time in the state of preservation of topographic features of certain wavelengths may signal a change in the thickness of the elastic lithosphere and thus in the amount of heat generated in and flowing from the interior.

A first-order distinction can be drawn between the planets on the basis of the continuity of their lithospheres. The moon, Mercury, and Mars, all with diameters half that of Earth or less, retain extensive regions of cratered terrain formed in the earliest part of their history (Fig. 1). Thus these smaller terrestrial planets, in contrast to Earth, show no evidence of the destruction and renewal of lithospheric plates over the latter 80 percent of their history. For these one-plate planets, thermal evolution helps to control both the state of stress in the lithosphere and the surface volcanic history. Net heating of the interior results in extensional stress in the lithosphere and is conducive to the ascent and eruption of magma. Net cooling causes global compression in the lithosphere and tends to inhibit the rise of magma to the surface.

The tectonic evolution of one-plate planets is dominated by several themes. The early formed global lithospheric plates underwent intense meteoroid bombardment. The largest impacts excavated cavities more than 1000 km in diameter and deeper than the thickness of the lithosphere. Relatively rapid sublithospheric flow may have extensively modified the cavity and the surrounding topography, leading to the development of basin ring structure. Major impact basins yielded topographic, thermal, and structural anomalies that geographically focused much of the subsequent tectonic and volcanic history of one-plate planets. Subsequent volcanic deposits collected in these low-lying regions, locally loading the lithosphere and often resulting in lithospheric flexure and failure.

A second major theme in the tectonic evolution of one-plate planets is the thickening of the lithosphere with time. Evidence for this is observed in the structure of ancient impact basins that have undergone extensive viscoelastic relaxation and in the response of the lithosphere to loads as a function of time. On the moon, for example, isostatic compensation of early loads by lithospheric deformation was nearly complete, while later loads have been partially supported for several billion years by a lithosphere of finite strength. Evidence for regional variations in lithospheric thickness at a given time is seen on the moon and Mars, and locally thinner regions are often the focus of extensive volcanic activity.

The influence of major changes in the interior and exterior environment of a planet are often readily visible on one-plate planets. While on Earth a change in the output of heat from the interior might be manifested as a change in plate recycling rates, on one-plate planets such a change could produce a global pattern of distinctive structures, such as the scarp system on Mercury. Changes in the direction of thermal evolution may also be recorded in the surface tectonics. On the moon, a change from extensional graben (rilles) in early mare deposits to compressional ridges in later units signals a change in the global state of stress in the lithosphere, reflecting a trend from net heating and expansion to net cooling and contraction.

The nature of the single lithospheric plate of smaller planets compared to the multiple, laterally mobile plates of Earth requires that tectonism on one-plate planets develop in a predominantly vertical sense. Vertical tectonism may be generally passive, caused by loading and depression of the lithosphere as in the case of the lunar mare basins, or it may be active, related to uplift over anomalous mantle, as has been suggested for the Tharsis bulge on Mars.

The larger Earth differs strikingly in the continuous relative movement of its numerous lithospheric plates, in the distinctive features associated with plate divergence and convergence, and in the large-scale recycling of lithosphere into the deeper mantle. A comparison of the ages of planetary surface units (Fig. 12) plainly shows that the majority of activity on the smaller terrestrial planets took place in the first half of the history of the solar system. Earth must have experienced many of the processes, such as impact basin formation, whose record is preserved on the smaller bodies, but the terrestrial record of such processes has largely been erased. The distinctive contrast in style and history might be attributed to the greater ratio of surface area to volume for the smaller planets than for Earth, leading to more rapid cooling and to the early formation and thickening of lithospheres. But what of Venus, similar in size to Earth? Preliminary topographic information (Fig. 11) shows some landforms unlike those on the smaller terrestrial planets and similar to such Earth features as the Tibetan Plateau. There is also evidence for impact craters, however, suggesting an extensive ancient surface on Venus.

Much of our ability to understand the most important processes affecting the tectonic evolution of the terrestrial planets rests on the future exploration of Venus. If higher resolution images show evidence of a plate tectonic origin for the rifts and mountain chains on Venus, then planetary size will remain the most plausible explanation for the differences in tectonic style among the planets. On the other hand, extensive ancient surfaces on Venus representing a single lithospheric plate, or evidence for a change from multiple plates to a single plate during the history of Venus, would indicate that other factors, such as temperature distribution or the presence of water, may be more critical than size in controlling the tectonic evolution of the terrestrial planets.

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A Decade of Progress in Earth's Internal Properties and Processes

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Summary. A major component of the Inter-Union Commission on Geodynamics Project, labeled "Internal Properties and Processes," included certain experimental and theoretical research in tectonophysics, seismology, geochemistry, petrology, volcanology, and planetology. This review focuses on a few research areas in which there have been surprises and reversals. In particular, attention is given to the attempts to quantify the thermal profile in the earth's interior and the material properties of the earth's interior.

One decade has passed since the Inter-Union Commission on Geodynamics (ICG) Project was launched, and 7 years ago the U.S. Program for the Geodynamics Project was published (*1*). A major component of this program, which has received national and international attention, was labeled "Internal Properties and Processes." It is appropriate to review the progress that has been made in this component of the ICG now that the Geodynamics Project has ended.

A comprehensive account of the progress in tectonics, seismology, and geochemistry appropriate to this component of the ICG is found in the U.S. National Report to the International Union of Geodesy and Geophysics, which is prepared every 4 years (*2*), and so I will not attempt a detailed review of that here. Instead, I will focus on a few research areas in which there have been surprises or in which there has been an overturning of the wisdom of the early 1970's.

The U.S. Program for the Geodynamics Project (*1*) listed several problems in "Internal Properties and Processes" that loomed in the early 1970's. The major headings were thermal structure, dynamical models, material properties in the deep interior, a theory of rheology, and the nature of instabilities in the deep interior. Of these problems, the thermal structure [in particular, the temperature (T) distribution of the earth] was selected as critical and worthy of accelerated and focused effort: "Temperature is probably the most important parameter concerning the state of motion of the earth's interior" (*1*, p. 32).

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The Thermal Regime

In 1970, future research on the thermal state required, or so it seemed, reliable ways of estimating the T profile. In other words, quantifying the unobservable internal T profile appeared to be an essential and feasible preliminary step in the task of interpreting the thermal regime.

That view is now no longer commonly held by those researching the thermal regime. Many now hold to the view that T is of secondary importance and that T has receded to the background as a "determined rather than a determining characteristic of the thermal regime of terrestrial planets" (*3*, p. 2). How this view changed is discussed below.

By 1970 many geophysicists no longer adhered to the long-entrenched belief that the thermal structure of a planet is controlled by heat conduction. Heat conduction theory imposed two insurmountable difficulties. First, the time (τ) required for a planetary-sized object formed from some initial T distribution to reach equilibrium is $\tau = R^2/k$, where R is the radius and k is the thermal conductivity. For the earth, τ is $\sim 10^{18}$ seconds, or longer than the age of the earth. Most of the original heat could not have dissipated, and for contemporary geodynamic processes a planetary-wide heat conduction theory means no significant heat flow from the interior while the process is occurring. The second difficulty is that the exact boundary conditions at the origin of the cooling cycle must be specified, and thus the earth's thermal regime is thereby tied closely to cosmology as well as to the radiogenic heat production (H).

A convection theory circumvented these difficulties. A substantial amount of heat is moved by mass motion in a convecting stream, and τ can be made small enough to be in consonance with