Large Igneous Provinces: A Planetary Perspective

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Large igneous provinces (LIPs) are common on the Moon, Mars and Venus, and their presence, characteristics, and geologic and temporal settings offer a potentially important perspective for interpreting LIPs on Earth. On the Moon, shallow magma reservoirs and large shield volcanoes are unknown. The relatively low-density, thick anorthositic crust creates a density trap for rising basaltic magmas which are thought to collect in reservoirs at the base of the crust; reservoir overpressurization causes dikes to propagate to the surface. Dikes sufficiently large to reach the surface are likely to result in large-volume, high-effusion-rate eruptions; single eruptive phases are predicted theoretically and observed in the maria to be several hundred to over $10^3$ km$^3$. On Mars, massive shield volcanoes have formed on the stable lithosphere over hot spots lasting over a billion years; shield heights are up to 25 km above the adjacent plains. Volumes of single edifices are of the order of $1.5 \times 10^6$ km$^3$, comparable to the total volumes of many basalt provinces on Earth. The impact cratering record on Venus suggests that Venus underwent rapid and massive planet-wide volcanic resurfacing about 300 m.y. ago, an event possibly related to the overturn of a depleted mantle layer resulting from the vertical accretion of a basaltic crust. This hypothesized event could be the equivalent of a planet-wide LIP and underlines the possibility of episodic and catastrophic LIPs throughout planetary history, resembling mantle overturn events proposed for Earth. The planetary record, in concert with the detailed examination of examples on Earth, can be of use in developing and testing models for the emplacement of LIPs, and may help to distinguish plate tectonic influences from those linked to deeper interior (mantle and core) processes.

INTRODUCTION AND BASIC CHARACTERISTICS OF TERRESTRIAL LARGE IGNEOUS PROVINCES

Recently, attention has been drawn to large igneous provinces on Earth, which are defined as regions of voluminous emplacement of predominantly mafic extrusive and intrusive rock whose origins lie in processes other than "normal" seafloor spreading [e.g., Coffin and Eldholm, 1992]. Large igneous provinces are characterized by transient large-scale intrusive and extrusive activity, including continental flood basalt (CFB) provinces (e.g., the Deccan Traps), volcanic passive margins (e.g., the Voring Margin), oceanic plateaus (e.g., the Ontong Java Plateau), ocean basin flood basalts (e.g., the Caribbean Flood Basalts), and large seamount chains (e.g., Hawaiian-
Emperor) [Coffin and Eldholm, 1992]. Commonly analyzed separately in the past, recent studies [e.g., Coffin and Eldholm, 1992, 1993] have shown that there are important temporal, spatial, and compositional relationships among terrestrial large igneous provinces, informally referred to as LIPS.

These studies, and numerous others that document individual occurrences (see references in this volume and those of Coffin and Eldholm [1994]), show that the genesis and evolution of LIPS are closely linked to mantle dynamics, that some LIPS represent major global events (large volumes of lava and associated intrusions are often produced in short episodes, which had potentially major effects on the global environment), and that emplacement of some LIPS may be related to changes in rate and direction of plate motion. Their formation may be episodic, but modification and destruction of older examples, and sedimentation and inaccessibility of others, makes this difficult to determine. Although several models have been proposed for the emplacement of LIPS (primarily associated with mantle plumes) [e.g., White and McKenzie, 1989; 1995; Griffths and Campbell, 1991; Larson, 1991a, b], these models are not yet well constrained by observations. At present only a limited, but growing (see articles in this volume), amount of quantitative data is available to assess associated mantle and crustal processes; to determine LIP dimensions, durations, rates of emplacement, crustal structure, and relationship to tectonism; and to predict environmental effects of LIP formation. For example, recent workers [Self et al., 1996, and this volume] have presented evidence that Columbia River flood basalt lavas may have been emplaced more gradually as inflating pahoehoe sheet flows forming very extensive flow fields rather than single very high-effusion-rate eruptions.

THE PERSPECTIVE FROM THE PLANETARY GEOLOGICAL RECORD

Large igneous provinces are also common on the terrestrial planetary bodies (Figure 1) other than the Earth [e.g., Basaltic Volcanism Study Project, 1981; Taylor, 1994], and their presence, characteristics, and geologic and temporal settings offer a potentially important perspective for understanding LIPS on Earth. For example, unlike the Earth, the majority of which is covered by water and thus virtually unknown at high spatial resolution, global imaging coverage exists for the solid surfaces of the Moon and Mars, and the Magellan project imaged over 98% of Venus at ~200 m resolution. In addition, exposure and preservation are excellent due primarily to fewer erosional agents, minimal erosional rates, and relatively stable lithospheres. Stable lithospheres also mean that longer time intervals are available for study (Figure 1). The age of over one-half of the Earth's surface (the ocean crust) is less than 5% of the age of the planet; the majority of the surface of the Moon and Mars, however, dates to the first half of solar system history. Terrestrial planetary bodies, by virtue of their number, offer multiple examples for study. Thus, LIPS might be studied in different places on one planet and among several planets. Similarly, the terrestrial planets provide an opportunity to assess how different environmental conditions (e.g., different crustal and thermal structure) might influence the formation and effects of LIPS. Furthermore, the segmented, laterally moving, and constantly renewing terrestrial lithosphere both insulates and obscures the view of many mantle convection processes and, indeed, is an active influence on these processes. One-plate planets [Solomon, 1977] such as the Moon, Mars, Mercury, and Venus can illustrate the long-term influences of mantle plumes and their variations under different thermal conditions in space and time. The multiple, well-exposed LIPS on the planets can also help to reveal the relation of plumes to tectonic structure. For example, Venus has tens of thousands of kilometers of exposed rift zones [Senske et al., 1992] which display a wide variety of igneous centers [Senske et al., 1992; Magee Roberts and Head, 1993], many of which are LIPS.

The planetary record can be instructive in terms of the chronology and episodicity of large igneous events and provinces. The extended historical record available for study (e.g., the first half of solar system history; Figure 1) permits an assessment of changes in the style of LIPS with time (potentially linked to thermal evolution, for example), and the frequency at any given time. Although radiometric dates from the planetary record are sparse, clues from well-exposed deposit morphology can sometimes even be used to estimate single-event duration [Head and Wilson, 1980]. Finally, the planetary record can offer a temporally complete perspective on many processes associated with LIPS. For example, lateral plate movement on Earth in the case of the Hawaiian-Emperor seamount chain (and other hotspot-related chains) helps to illustrate many stages in hotspot development by spreading the signature out into a series of volcanic edifices; this same process, however, destroys the signature of the initial plume which presumably has been subducted under Kamchatka. On the planets, particularly Venus, the start-to-finish processes of mantle plumes can be studied (e.g., the relation of thermal uplift, tectonics, and volcanism in a single example and from examples in different stages of formation) [e.g., Stofan et al., 1992; Keddie and Head, 1994a] and compared to Earth. In summary, the planetary record, in concert with detailed examination of examples on Earth, should help to
develop and test models for the emplacement of LIPs, and
to distinguish plate tectonic influences from those linked to
deeper mantle and core processes.

Terrestrial planetary bodies show a wide variety of charac-
teristics: e.g., size, density, gravity, presence/absence and
nature of atmospheres, thermal evolution, and starting
conditions (Figure 1). Obviously, all comparisons to the
terrestrial record must keep these variations and dif-
fences in mind, as well as the positive aspects of
comparative planetology described above. The planetary
record is not, of course, a panacea. In many cases
available information for specific aspects of different
planets (e.g., the detailed crustal thickness and structure on
Mars and Venus) is limited, resulting in some uncertainties
involving correlations, relationships, and causal factors.
Nonetheless, the information provided by specific
examples and the perspective provided by considering
different conditions on different planets should contribute
to our understanding of the formation and evolution of
LIPs. The purpose of this paper is to present a range of
specific examples and to explore the potential application
of these examples to current problems in understanding
LIPs on Earth.

**PLANETARY EXAMPLES**

We proceed in order of increasing planetary size (Figure
1), first describing the general crustal, lithospheric, tec-
tonic, and temporal setting of basaltic volcanism on the
Moon, Mercury, Mars, and Venus, and then discussing
specific examples of large-volume basaltic magmatism on
each planetary body and the potential relevance to LIPs on
Earth.

The Moon

The Moon's diameter is about one-quarter that of the
Earth. The Moon is of lower density, has not retained an
atmosphere, is characterized by vertical tectonics of an
unsegmented lithosphere (not lateral plate tectonics), and
now has a very thick lithosphere. Most of its geological
surface activity took place in the first half of solar system
history (Figure 1) [e.g., Head and Solomon, 1981]. Informa-
tion about the Moon comes from remote observations and
surface exploration, including returned samples and
scientific stations [see Heiken et al., 1991].

Basaltic volcanic deposits on the Moon consist largely of
lunar maria which cover about 17% of the surface,
primarily on the near side (Figure 2a). The total area of
the lunar maria ($6.3 \times 10^5 \text{ km}^2$ [Head, 1975b; 1976]) is
considerably larger than typical terrestrial LIPs but only
slightly larger than the area of the Ontong Java LIP (Figure
3). The lunar maria were emplaced over about 1.5-2
billion years, largely in the first half of solar system history [Wilhelms, 1987], but the total volume was relatively small, about $1 \times 10^7$ km$^3$ [Head, 1975b]. This value for the total planet is comparable to the volume of the Deccan flood basalt deposits alone, but considerably less than the total present volume of the terrestrial oceanic crust, about $1.7 \times 10^8$ km$^3$ (Figure 3). The average lunar global magma flux was low, about $10^{-2}$ km$^3$/a, even at peak periods of mare emplacement (in the Imbrian Period, 3.8–3.2 Ga). This average global flux is comparable to the present local output rates for such individual terrestrial volcanoes as Kilauea or Vesuvius. Output rates for individual eruptions on the Moon were occasionally extremely high; several individual eruptions associated with sinuous rilles may have emplaced more than $10^7$ km$^3$ of lava in about a year [Hulme, 1973], a single event that would represent the equivalent of about 70,000 years of the average flux!

Volcanic features manifesting large-volume eruptions include the individual maria themselves, extensive flow fronts, some stretching for distances of over 1200 km [Schaber, 1973], volcanic complexes that might signal the location of hotspots, and sinuous rilles, which have been attributed to high-effusion-rate eruptions involving thermal erosion of the substrate. Interestingly, no large shield volcanoes, such as those seen on the Earth (e.g., Hawaii), Mars (e.g., Olympus Mons), or Venus (e.g., Sapas Mons) are observed on the Moon; large caldera-like features are also extremely rare [Head and Wilson, 1991].

The lunar maria are of diverse sizes and shapes [Head, 1975a], and individual mare occurrences might be thought of as equivalent to some terrestrial LIPs (Figure 3), particularly those maria that tend to be concentrated within large impact basins of various states of preservation [Head, 1975a, b; 1976]. Indeed, Alt et al. [1988] proposed that
flood basalts that form within plates, with no apparent tectonic cause, are the terrestrial equivalents of the lunar maria. In their model, an impact crater on Earth large enough to cause pressure-release melting would be quickly flooded to form a lava lake (equivalent to the lunar maria) and these events, in turn, would initiate hotspots, which would develop into persistent low-pressure cells within the mantle [Alt et al., 1988]. Does the lunar record support this model? Although early theories suggested a causal relationship between lunar impact basin formation and basaltic mare filling, the results of the Apollo and Luna exploration programs and models of basin formation and evolution [Solomon et al., 1982; Bratt et al., 1985] showed that generation of basalts via impacts was unlikely and that
impact basin formation and filling by mare basalt are separated in time. In the case of the 900-km-diameter Orientale impact basin, vast quantities of substrate were impact-melted by the basin-forming event to produce a sheet of high-albedo plains lining the basin interior and floor, and are estimated to have a volume of ~200,000 km$^3$ [Head, 1974]. This unit has a compositional affinity to the non-mare target rocks [Head et al., 1993] and is distinctly different in composition and age from the adjacent basaltic maria deposits, which span an interval of several hundred million years [Greeley et al., 1993]. In most other mare basins, the vast majority of the exposed volcanic plains were emplaced over several hundred million years following the impact event [Basaltic Volcanism Study Project, 1981]. There is no evidence for the production of basin-sized lunar basaltic “lava lakes which crystallized from the surface down” [Alt et al., 1988]. The stratigraphy of lunar maria infilling documents both the long and sequential development of extrusive events, and the difference in age between the basin-forming event and its basaltic lava filling. Localization of the maria in the basins apparently was due to passive variations in crustal thickness and ponding in topographic lows [Head and Wilson, 1992a], processes discussed further below.

Although the equivalence of an impact origin of a basin and its fill on the Moon and LIPs on Earth proposed by Alt et al. [1988] is not supported by evidence from the Moon, impacts on Earth could potentially initiate volcanism. The small size of the Moon (and correspondingly very different pressure gradient), its thicker crust, and its variable lithospheric thickness could all inhibit melting relative to a comparable event on Earth. Convincing arguments have been put forth to indicate that impact-initiated volcanism was not a factor in the large (~200 km diameter) Sudbury basin formed in continental crust on Earth [Grieve et al., 1991a]. Similar-size impacts into thin crust and lithosphere typical of a young oceanic floor setting could conceivably produce pressure-release melting and associated volcanism [Rogers, 1982]. Craters typically formed during the time of emplacement of most well-documented LIPs (e.g., the last 250 m.y. [Coffin and Eldholm, 1994]) are characterized by relatively small size, shallow depths of excavation, and lack of significant lava fill [Grieve et al., 1991b; but see also Oberbeck et al., 1993]. Large-scale riftng and deep-source plume volcanism are more likely candidates for LIP formation and evolution during this time period. In early Earth history, however, very large impacts into ocean crust and thin lithosphere may have been sites of extensive volcanism caused by mantle uplift and decompressional melting [e.g., Grieve, 1980; Frey, 1980; Grieve and Parmentier, 1984].

Other large volcanic accumulations on the Moon include the extensive lava flow fronts of Mare Imbrium which were emplaced at least a billion years after the formation of the impact basin. These occur in three phases which extend 1200, 600, and 400 km from the southwestern edge of the basin into its interior. The three flow units have a total volume of >4 × 10$^8$ km$^3$, and very high effusion rates are implied by their lengths and volumes; effusion rates and flow volumes are comparable to some of those reported for the Columbia River flood basalts [Schaber, 1973; Tolan et al., 1989], although fractal analyses raise the possibility that the Imbrium flows could have been emplaced as numerous thin pahoehoe flows [Bruno et al., 1992]. The fact that these units are some of the youngest on the Moon suggests that other more degraded flows filling the earlier lunar maria may also have been emplaced similarly. Examination of isolated mare basalt ponds in the highlands fringing the continuous maria has shown that typical volumes range from 100 to 1200 km$^3$, values similar to those of terrestrial flood basalts eruption units [Yngst and Head, 1994; 1995; Tolan et al., 1989]. Thus, many of the individual eruptions that make up the maria may be equivalent to units within flood basalts and LIPs on Earth, but the eruption frequency seems to have been much less; the lunar maria were emplaced over many hundreds of millions of years, rather than a few million years as was apparently the case in most terrestrial examples.

Another unusual characteristic of lunar maria relative to LIPs on Earth is sinuous rilles (Figure 2c), which are meandering channels preferentially located along the edges of the maria [Schubert et al., 1970]. They range up to about 3 km wide and from a few kilometers to more than 300 km long. Sinuous rilles are generally an order of magnitude larger and often much more sinuous than terrestrial lava channels. Many characteristics of lunar sinuous rilles unexplained by simple lava channel, tube or other models [e.g., Oberbeck et al., 1969, Grieve, 1971; Spudis et al., 1987] can be accounted for by thermal erosion [Hulme, 1973, 1982; Carr, 1974]. The length, width, and depth of large sinuous rilles and the nature of their source regions provide important information on eruption conditions. For a 50-km-long rille in the Marius Hills, Hulme [1973] calculated an effusion rate of 4 × 10$^4$ m$^3$/s, an eruption duration of about one year, and a total magma volume of about 1200 km$^3$. The sizes of source depressions of sinuous rilles provide independent evidence for extremely high-effusion-rate eruptions of long duration [Wilson and Head, 1980; Head and Wilson, 1980]. On the basis of these studies, key factors in the formation of sinuous rilles by thermal erosion are (1) turbulent flow, requiring high effusion rates and aided by low yield strength and (2) sus-
tained flow (implying very long-duration eruptions and thus very high eruption volumes) to cause the continued downcutting of the rille to the observed depths. Thus, eruptions that caused many of the large sinuous rilles on the Moon were apparently characterized by rapid effusion of low-yield-strength lavas for prolonged periods, producing flows of extremely high volumes (in the range 300–1200 km$^3$), comparable to those in terrestrial flood basalt provinces (e.g., the ~1375 km$^3$ Roza Member of the Columbia River Basalt [Martin, 1989]). In contrast, typical eruption volumes for shield-related flows on Earth are much less than a cubic kilometer [Peterson and Moore, 1987], with the largest historic lava flow (Laki) being about 15 km$^3$ [Jonsson, 1983; Thordarson and Self, 1993].

Several mare-related areas show unusual concentrations of volcanic features on the Moon [Guest, 1971; Whitford-Stark and Head, 1977]. Two of the most significant of these (Figure 3) are the Marius Hills area (35,000 km$^2$), which displays 20 sinuous rilles and over 100 domes and cones, and the Aristarchus Plateau/Rima Prinz region (40,000 km$^2$) which is dominated by 36 sinuous rilles (Figure 2c). The high concentration of sinuous rilles suggests that these complexes are the sites of multiple high-effusion-rate, high-volume eruptions and that these centers may be the surface manifestation of hotspots [Head and Wilson, 1992a] and thus possible analogs to terrestrial LIPs. The thick crust (about 60–80 km) and lithosphere (in excess of the thickness of the crust) characteristic of the Moon (and thus the greater depths of magma sources) may make these candidate hotspots less recognizable and more analogous to continental volcanic provinces. In addition, lava flow deposits on the Moon are much more widely dispersed from their sources.

In summary, the lunar maria are comparable in scale to some terrestrial LIPs (Figure 3) but on the basis of available data appear to have been emplaced over much longer periods of time (e.g., $10^8$ to $10^9$ years rather than $10^6$ to $10^7$ years). Many individual eruptions, however, appear to be similar in volume and eruption rates to those in flood basalt provinces [Tolan et al., 1989]. Little evidence exists for shallow magma reservoirs and repeated small-volume eruptions that would build up large shield volcanoes. The observed characteristics seem to call for large batches of magma erupted over short periods of time from relatively deep sources but separated in time by significant intervals. How can these characteristics be accounted for in terms of the nature of the source regions and the modes of emplacement?

One model [Head and Wilson, 1992a] begins with the observations that the basaltic maria are superposed on the ancient, globally continuous, and thick low-density anor- thositic highland crust, the latter derived primarily from global-scale melting associated with planetary accretion. The low-density highland crust provided a density barrier [Solomon, 1975] to ascending mantle plumes and basaltic melts. In this view, rising diapirs and magma bodies tended to collect at the base of the 60–80 km thick crust (Figure 2b). Following sufficient overpressurization of source regions by partial melting or arrival of additional material into the reservoir, individual dikes propagated toward the surface. Thus, the thick highland crust created a deep zone of neutral buoyancy for rising magma that could only be overcome by overpressurization events which caused dikes to propagate to the surface.

In this model, whether intrusion or eruption occurred was determined by variations in overpressurization and crustal thickness. Low levels of overpressurization resulted in intrusion into the lower crust, forming dikes which cooled and solidified. Dikes characterized by sufficient overpressurization to approach the surface could have several fates. Overpressurization events large enough to propagate dikes to the surface to cause eruptions are predicted to involve very large volumes of magma [Head and Wilson, 1992a], comparable to those associated with many observed lava flows, such as the flows extending hundreds of kilometers into Mare Imbrium [Schaber, 1973] and those associated with sinuous rilles. Intrusion close enough to the surface to produce a distinctive near-surface stress field often resulted in the production of linear graben-like features along the strike of the dike and small associated effusions and eruptions. In the case of the linear graben Rima Parry V, small spatter cones are aligned along the central part of the graben [Head and Wilson, 1994a]. Dikes propagating to slightly deeper levels may not create near-surface stress fields sufficient to form graben, but subsequent degassing may form chains of pit craters over the site of the dike.

The model predicts that the relationship between the size of the magma source and highland crustal thickness was such that dikes propagated to the near-surface and surface relatively infrequently (Figure 2b). Thus, most dikes had sufficient time to cool before the next dike was emplaced. Frequent emplacement of dikes to create a shallow magma reservoir was very difficult on the Moon. The lack of Hawaii-like shield volcanoes and the paucity of caldera-like features are thus attributed to the difficulty in producing shallow magma reservoirs which result in emplacement of many individual flows, edifice-forming flows, and associated calderas [Head and Wilson, 1991]. In addition, the same lack of multiple, continuous dike emplacement events of sufficient magnitude to reach the surface over short periods of time meant that the lunar
maria tended to be produced from relatively large eruption events spaced over very long intervals, in contrast to terrestrial LIPs.

The lunar situation described in this model is analogous in many ways to basaltic magma bodies interacting with terrestrial continental crust. On Earth, zones of neutral buoyancy [e.g., Glazner and Ussler, 1988] stall buoyantly rising basaltic magma bodies within the crust. Overpressurization events can cause the same features seen on the Moon, as exemplified by many of the basaltic volcanic fields in the western United States [e.g., Crumpler et al., 1994], and indeed large-scale flood basalts can be emplaced that are comparable in size to the large lunar flows [Tolan et al., 1989]. The low melting temperature of the continental crust relative to that of the more refractory lunar anorthosic crust means that stalled basaltic magma bodies in continental crust may cause associated and large-scale crustal melting, resulting in a geochemical and petrologic complexity unknown on the Moon. The continental crust and the lunar highlands illustrate the role of large-scale density barriers impeding the creation of significant shallow basaltic reservoirs, such as those observed at seafloor spreading centers and in large edifices such as Hawaii. Complex shallow reservoirs do exist in continental crust, however, where local conditions of melt generation and, unlike on the Moon, sustained supply rates exist (as in continent margin subduction zones and hotspot traces or rifting environments). In these cases, composite volcanoes are common. No known analog of these features exists on the Moon and Venus, but several examples may be present on Mars (e.g., Hecates Tholus [Mouginis-Mark et al., 1982; Wilson and Head, 1994; Hodges and Moore, 1994]).

Mercury

Mercury remains one of the most enigmatic and promising planets in the inner solar system in terms of understanding the relationship of its unusual interior to its volcanic and magmatic history [Chapman, 1988]. Information about Mercury comes from the Mariner 10 mission and Earth-based observations [see Vilas et al., 1988]. Mercury is about one-third the diameter but approximately the same density as the Earth, has not retained an atmosphere, and is characterized by vertical and some lateral tectonics of a largely unsegmented lithosphere (not lateral plate tectonics). Most geological surface activity took place in the first third of solar system history (Figure 1) [Head and Solomon, 1981; Vilas et al., 1988]. The very high density of Mercury relative to its size has been attributed both to initial temperature-pressure conditions in the inner part of the condensing solar nebula, which favored retention of refractory components [e.g., Goettel, 1988], and to the effects of a giant impact event stripping off a low density crust and upper mantle after core formation [Cameron et al., 1988].

Mercury is poorly explored in terms of photographic coverage and remote sensing data [Chapman, 1988]. Knowledge of internal structure is meager, although a high-density core comprising well over one-half Mercury’s diameter (about the size of the Earth’s Moon) is likely. In addition, prominent albedo variations such as those that distinguish the lunar maria from the heavily cratered highlands are not apparent on Mercury. Smooth plains are present, but a possible volcanic origin cannot readily be distinguished from plains produced by ponding of impact ejecta, a process known to occur in the light plains surrounding impact basins on the Moon [Oberbeck, 1975; Oberbeck et al., 1975; Wilhelms, 1976]. The stratigraphy and geologic history of Mercury suggest that major volcanic provinces were emplaced in the first third of solar system history [Spudis and Guest, 1988], but the details are insufficient to provide a basic characterization of such provinces or an understanding of their mode of emplacement. If these plains are indeed of volcanic origin, their general lack of associated volcanic features [Strom et al., 1975; Trask and Strom, 1976] suggests possible flood basalt emplacement.

Mars

Information about Mars [e.g., Kieffer et al., 1992a] comes from Earth-based observations, extensive spacecraft exploration (including orbiters and landers [e.g., Kieffer et al., 1992b; Snyder and Moroz, 1992]), and meteorites believed to be ejected from Mars by impacts and transported to Earth [e.g., Longhi et al., 1992]. Mars is about one-half the diameter and of much lower density than the Earth, has a thin CO₂ atmosphere, and is characterized by vertical tectonics of an unsegmented lithosphere (not lateral plate tectonics); most of its major geological surface activity took place in the first half of solar system history, with some volcanism and significant eolian activity continuing well into the last half of solar system history (Figure 1, 4a) [e.g., Head and Solomon, 1981; Kieffer et al., 1992b]. The total area of Mars covered by volcanic material has been estimated to be about 58% of the surface (≈0.84 × 10⁶ km²) [Tanaka et al., 1988], and the total volume of surface extrusion to be 2 × 10⁶ km³ [Greeley, 1987] (Figure 3). The corresponding intrusive volume is not known but is likely to be larger by at least a factor of 10, the ratio typical of the continental
regions on Earth [e.g., Crisp, 1984; see Wilson and Head, 1994]. Volcanism has decreased over geologic time from broad regional resurfacing to local activity; areal resurfacing rates have steadily decreased from ~1 km² a⁻¹ to ~10⁻² km² a⁻¹ [Tanaka et al., 1992].

On Mars, in contrast to the Moon, large shield volcanoes have been emplaced that resemble those on Venus and the Earth in morphology. They exhibit a wide range of rift zone development, internal deformation related to lithospheric loading and flexure, flank and slope failure, and summit caldera development [Carr, 1973, 1981; Hodges and Moore, 1994; Wilson and Head, 1994; Crumpler et al., 1996]. Their scales are quite different, however (Figure 4b, c). Martian shields possess breadths of many hundreds of kilometers, and their heights are commonly a factor of three greater than Hawaii (up to 25 km!). Volumes of individual shields are gigantic (Figure 3). Olympus Mons (Figure 4c) has a volume of about 2 × 10⁶ km³ (above its base), compared to 1 × 10⁵ km³ (above its base) for the island of Hawaii (which is composed of several different shields) and 1.1 × 10⁶ km³ for the whole Hawaiian-Emperor seamount chain [Barger and Jackson, 1974]. Volumes of other single edifices are of the order of 1.5 × 10⁶ km³, comparable to extrusive volumes estimated for the Karoo, Paraná, Deccan and North Atlantic basalt provinces on Earth (Figure 3). Martian caldera structures

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**Figure 4.** Mars. (a) Generalized topography and geography. The dashed line extending from upper left (near Arcadia Planitia) to lower right (near Apollinaris Patera) across both hemispheres separates the northern lowlands from the southern highlands; other closed dashed lines are impact basin depressions (e.g., Argyre Planitia, Hellas Planitia) or volcanic provinces (e.g., Hesperia Planum; Lunae Planum). Black spots are shield volcanoes of various sizes; the concentration of shield volcanoes in the left hemisphere is the Tharsis region (see Figure 4b for enlargement) and in the right hemisphere is the Elysium region. (b) Geologic sketch map of the Tharsis region. Topography is indicated by contour lines at 5 km intervals with tick marks pointing downslope. Stars mark summits of the major shield volcanoes (Figure 4a), Olympus Mons and the Tharsis Montes (Arsia, Pavonis, and Ascraeus), which commonly reach elevations in excess of 20 km; upper contours are omitted. Width of diagram is about 4000 km at the equator and the units are discussed in the text. Map is from Head and Solomon [1981] from data of Wise et al. [1979] and Scott and Carr [1978]. (c) Oblique view of Olympus Mons volcano, one of the large lava shields in the Tharsis region (Figure 4a, b); summit is about 25 km above the base of the volcano and is characterized by a complex caldera and two nearby impact craters. Flows emanating from near the summit extend down the flanks and often cascade over the several-kilometer-high scarp at the base of the volcano. Viking Orbiter photograph VO 641A52. (d) Stratigraphic sequence showing context and main events in the evolution of Tharsis in relation to global processes [from Banerdt et al., 1992]. Locations of regions are shown in Figure 4a, b. Absolute ages are from the time-scale models of Hartmann-Tanaka (H/T) [Hartmann, 1978] and Neukum-Wise (N/W) [Neukum and Wise, 1976], as summarized by Tanaka et al. [1992].
in longer cooling-limited flows and wider dikes characterized by higher effusion rates [Wilson and Head, 1994]. Because the lithosphere has been stable and has not moved laterally over the majority of martian history, regions of melting in the mantle (e.g., mantle plumes) concentrate their effusive products in a single area, rather than having them spread out in conveyor-belt-like fashion, as in the case of the Hawaiian-Emperor seamount chain on the Pacific Ocean floor. Thus, melt products accrete vertically into huge accumulations [Carr, 1973], loading the lithosphere and causing flexure, deformation, and edifice flank failure.

The extreme height of martian volcanoes also appears to be related to lithospheric structure. Comer et al. [1985] examined deformational structures surrounding several Tharsis-region volcanoes (shown as large black spots in Figure 4a) to assess lithospheric flexure caused by volcano loading and to estimate the thickness of the elastic lithosphere. They found that elastic lithosphere thicknesses are in the range of 20–50 km for regions surrounding the majority of the Tharsis shields. The lithosphere appears to be at least 150 km thick in the region of Olympus Mons. Thus, one factor contributing to the large height of the martian volcanoes is the relatively thick elastic lithosphere during their formation; the volcanic load and underlying lithosphere did not subside at a rate that would limit their heights. In addition, variations in lithospheric thickness in

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Figure 4. (continued)
space and time can be very important in the construction and subsequent modification of volcanic edifices. For the volcanoes forming the Galápagos Archipelago, Feighner and Richards [1994] showed that lithospheric thickness is related to volcano size and structure across the archipelago between areas of effective elastic lithospheric thickness of 6 and 12 km. McNutt et al. [1989] demonstrated that the thermal and mechanical state of the lithosphere apparently controls the expression of weak plumes such as the Marquesas. In a study relevant to terrestrial shield volcanoes and to the history of lava emplacement in LIPs, McGovern and Solomon [1993] modeled lithospheric flexure and time-dependent stress and faulting on the Tharsis volcanoes and demonstrated sufficiently large flexural stresses in several examples to cause failure by faulting. Such stresses in turn could have influenced the subsequent path of magma ascent and emplacement.

One of the most impressive global-scale features on Mars is the Tharsis region, a LIP comprising ~20% of the surface area of Mars that dwarfs those presently known on Earth in size, associated features, and duration (Figure 3). Tharsis, which forms a broad dome or rise about 4000 km in diameter rising as much as 10 km above surrounding terrain, dominates the western hemisphere of Mars (Figure 4b). Its area of ~6.5 × 10^6 km^2 is larger than the largest known terrestrial LIP and totals over one-half the total area of the lunar maria (Figure 3). The Tharsis rise is composed of areally extensive volcanic plains spanning a wide range of ages; massive superposed shield volcanoes (e.g., >500 km wide and up to 25 km high) are associated with tectonic features that include radial fractures and graben extending beyond the rise and perhaps associated with uplift, and concentric wrinkle ridges indicating crustal shortening. The volcanic deposits clearly associated with this province, the western volcanic assemblage [Tanaka et al., 1992], cover an area of 1.4 × 10^7 km^2 (Figure 3); more-degraded deposits may also be volcanic.

On the basis of geologic mapping at a variety of scales [Scott and Tanaka, 1986, and summarized by Tanaka et al., 1992] a general stratigraphy and chronology for Tharsis has begun to emerge (Figure 4d). In contrast to many terrestrial LIPs which formed over 10^5–10^6 years, these data point to volcanic and tectonic activity in the Tharsis region spanning 10^8–10^9 years. Ancient cratered terrain bounds Tharsis to the south and is exposed at high elevations within Tharsis, suggesting extensive uplift. Plains units interpreted as volcanic and major shield volcanoes dominate the rest of Tharsis. Undivided plains (pu in Figure 4b) make up the vast majority of surface units and extend north of Tharsis. Ridged plains (pr) are characterized by many mare-ridge type features indicating shortening. Major volcanic edifices and structures dot the surface of Tharsis and four of these (indicated by stars on their summits in Figure 4b) exceed 25 km in elevation above the surrounding terrain (Figure 4c). Younger volcanic plains units (pt) surround the major central shield volcanoes, their most likely source. Tectonic features are abundant and the most prominent of these, the Valles Marineris rift system, extends several thousand kilometers from central Tharsis toward the east. Its floor is indicated by unit cf, canyon floor materials, in Figure 4b.

Tharsis rise development involved complex episodic tectonism and intimately associated volcanism on both local and regional scales. Early fractured plains (Noachian and Hesperian Epochs, Figure 4d) made up mostly of volcanic rocks erupted during the early stages of Tharsis activity are cut by the most intense deformation in Tharsis, represented by fault systems that are radial and concentric to volcanic centers such as Tharsis Montes and Alba Patera. These faults formed during the Noachian and Hesperian epochs; concentric ridge systems representing local shortening were formed mainly at distances greater than 2000 km from the center of the Tharsis rise, primarily during the Late Noachian and Early Hesperian Epochs. The latest faulting occurred in the Amazonian Epoch (Figure 4b, d) primarily in association with the active volcanic centers mentioned above [Tanaka et al., 1992]. The Tharsis Montes (from south to north, Arsia, Pavonis and Ascraeus Montes; Figure 4a, b), composed of three massive shield volcanoes aligned in a row along the crest of the Tharsis rise, are the primary sources for the volcanic Tharsis Montes Formation (largely unit pt in Figure 4b), which covers an area of almost 7 × 10^6 km^2 and is composed of lobate sheet flows, some of which extend almost 1500 km from the source shields [Schaber et al., 1978; Plescia and Saunders, 1979]. Olympus Mons, a similar shield volcano to the west of Tharsis Montes, is the source area for Upper Amazonian lava flows, some of the youngest on Mars (Figure 4b, c).

Theories to account for the Tharsis rise abound [see discussions by Schubert et al., 1992; Banerdt et al., 1992; and Tanaka et al., 1992]. Initial ideas centered on an area of convective upwelling producing a very large mantle plume which generated uplift and volcanism [e.g., Carr, 1974], an idea supported by calculations of mantle convection under martian conditions in which a limited number of convection cells are favored [e.g., Schubert et al., 1990]. The interpreted topographic uplift, however, could not be explained by these dynamic processes alone. Isostatic uplift caused by lateral migration and intrusion of material thermally eroded from the base of the crust of the northern lowlands was favored by Wise et al. [1979].
whereas Schultz et al. [1982] suggested preferential concentration of volcanism along early impact-basin ring structures. Debate also centers on the relative role of volcanism and uplift, with some workers preferring major uplift and relatively minor volcanism [e.g., Plescia and Saunders, 1980] and others suggesting that Tharsis resulted from an extended period of regional volcanism in an area of thin lithosphere [e.g., Solomon and Head, 1982]. The inability of stress models to account simply for the extensive radial graben systems has led workers to accept the idea that more than one mechanism of lithospheric deformation is required; simple isostatic or flexural loading models do not satisfy all observations. The present conundrum is that stress models seem to require two different events, but the geologic evidence suggests that the radial graben formed essentially simultaneously [e.g., Banerdt et al., 1992].

Finnerty et al. [1988] constructed a quantitative petrologic model for Tharsis which was extended to a more general model for the evolution of Tharsis [Phillips et al., 1990]. Using melt partitioning data and models for likely martian mantle compositions, they showed that extraction of basalt melt from the mantle and subsequent crustal intrusion and extrusion could have resulted in a net volume increase in the crust-mantle column, producing a prominent topographic rise with no net increase in mass. Much of the support for the uplift would come from the source-region residuum, and most of the magma produced by the required melting must end up as intrusions in the crust and upper mantle. Although consistent with many of the major characteristics of Tharsis, these models do not easily satisfy the gravity data.

What currently supports the Tharsis region, some several billion years after its initial activity? Gravity data show an extremely large free-air anomaly [e.g., Esposito et al., 1992]; simple isostatic compensation is essentially ruled out and dynamic support by active mantle flow is very unlikely because of the difficulty of maintaining such large-scale and consistent mantle flow for several billion years. Many models have been proposed, and the most likely have the Tharsis rise partially supported by the elastic strength of the lithosphere, with additional support from the buoyancy of a crustal root at depths of about 50–100 km [e.g., Banerdt et al., 1992]. The early history of the Tharsis rise might have involved a transient mode of support (e.g., a convective plume, or an upper mantle density-deficit induced by thermal or chemical factors) and a regional crustal thickness about 25–30 km in excess of that estimated in global-scale models. Subsequent reduction and removal of the transient support were accompanied by the general cooling of the planet, leaving a superisostatic load on a cooling, thickening lithosphere.

A second large domal area on Mars, the Elysium rise, is about 2000 km across (Figure 4a). Although smaller than Tharsis, it also has high concentrations of volcanism and tectonism [Greeley and Guest, 1987; Mougins-Mark et al., 1984], including several volcanic formations, and three major shield volcanoes (Albor Tholus, Hecates Tholus, and Elysium Mons). Hall et al. [1986] argued on the basis of thermal and mechanical arguments that flexural uplift preceded or was contemporaneous with the emplacement of the majority of the volcanic deposits.

Why does Mars have two prominent, long-lasting, extremely large igneous provinces? Convective plumes in the martian mantle were modeled using numerical simulations of fully three-dimensional convection in a spherical shell [e.g., Schubert et al., 1990, 1992]. These models suggest that cylindrical plumes are the most probable form of upwelling in the mantle and that downwelling occurs in an interconnected network of planar sheets; the number of upwelling plumes is a function of the geometry of heating. Increase in bottom heating causes a decrease in the number of upwellings and an increase in their intensity, with very substantial bottom heating producing only six plumes. Gradual cooling of the planet, and the core in particular, means that the planform and style of convection likely changed with time, with fewer, more vigorous plumes earlier and more, smaller plumes later. In addition, the temperature-dependence of mantle viscosity will have an influence on plume structure and abundance. Although the trend in early history might have been toward a small number of vigorous plumes, a variable lithospheric thickness and a thickening lithosphere with time [e.g., Comer et al., 1985; Solomon and Head, 1990] might hide the surface effects of all but the most prominent plumes.

What are some possible lessons for those who study LIPs on Earth? First, it is clear that LIPs can achieve massive proportions and form over long periods. Tharsis covers 20% of the surface area of the planet Mars and was active for several billions of years. In addition, the martian LIPs confirm that scale and total duration of igneous emplacement can change as a function of time and thermal evolution (large-scale planetary cooling). Early plumes might have been less numerous, larger, and more vigorous due to a larger role of bottom heating. Planetary thermal evolution will also influence lithospheric thickness and the surface manifestation of plume impingement; thus we should anticipate considerable variability in LIPs through time. The abundance of large shield volcanoes within Tharsis, each of which would qualify as a LIP on Earth (Figure 3), also suggests that individual plumes are likely within a larger diffuse upwelling such as may have formed Tharsis as a whole. Given the incompleteness of the terrestrial record, the martian record suggests that some
single LIPs on Earth might be only one “tree” in a larger “forest” of a megaplume. Finally, the petrogenetic effects of shallow melting and the resulting residuum might leave depleted mantle signatures that could persist for hundreds of millions to billions of years, even on a planet as dynamic as Earth.

Venus

Venus is approximately the same diameter and density as the Earth and is Earth’s closest planetary neighbor (Figure 1). These similarities have led to frequent comparison of Venus with the Earth and the idea that Venus might be a “sibling” or possibly even a “twin.” Venus offers an important test of major ideas about planetary evolution in terms of the role of planetary size and initial position in the solar system [Head and Solomon, 1981], crustal formation and evolution [Head, 1990b], and mechanisms of lithospheric heat transfer [Solomon and Head, 1982]. Although Venus has many similarities with the Earth, it also has important differences. It has a thick, dense CO₂ atmosphere, rotates very slowly, and has essentially no magnetic field. The high surface pressure is approximately comparable to that on the ocean floor below about a kilometer depth, and average surface temperatures are around 475°C, precluding the presence of liquid water and resulting in the preservation of landforms in their near-pristine state. Information on the nature of Venus has come from Earth-based observations and a variety of planetary probes, including flybys, orbiters, balloons, atmospheric probes and landers [e.g., Hunten et al., 1983; Cruikshank, 1983]. Data from the recent Magellan mission provided high resolution radar image coverage for almost the whole planet and altimetry and gravity data [Saunders et al., 1992]; this information, together with data from previous US and Soviet missions, has resulted in a more comprehensive view of the geology and geophysics of Venus (Figure 5a).

Approximately 80% of the surface of Venus is made up of volcanic plains and a wide variety of volcanic landforms (Figure 5b-e), probably largely basaltic in composition [Head et al., 1992]. Unlike on the Earth, volcanic landforms are not distributed along elongated plate boundaries and hotspot traces; rather, they are broadly distributed over the whole planet and also clustered in a large region (Beta-Atla-Themis) making up about 20% of the surface [Crumpler et al., 1993]. Although Venus has folded mountain belts [Crumpler et al., 1986], global rift zones [Senske et al., 1992], and features that resemble Earth’s convergent plate boundaries [Head, 1990a; McKenzie et al., 1992], Magellan revealed no evidence for the extensive global plate-tectonic boundaries and crustal structures that would indicate ongoing crustal spreading and recycling [Solomon and Head, 1982; 1991; Solomon et al., 1992].

The ~80% of the surface area of Venus estimated to be covered by volcanic plains (~3.68 × 10⁶ km² [Head et al., 1992]) can be combined with an estimate of the average plains thickness of about 2.5 km based on stratigraphic relationships [Head et al., 1996a] to predict the total volume of surface extrusion of about 9.2 × 10⁶ km³ (Figure 3). On the basis of impact crater counts on volcanic units, volcanism has apparently decreased over geologic time from a period of global resurfacing to much less voluminous local activity, with average effusion rates changing from about 5 km³/a to <1 km³/a [Head et al., 1992].

On the Earth, typical basaltic melts are positively buoyant, but can stall at a neutral buoyancy zone (NBZ) representing a near-surface, low-density horizon related to weathering and gas-exsolution porosity [e.g., Ryan, 1988]. The results of theoretical modeling indicate that very high atmospheric pressure on Venus reduces volatile exsolution and magma fragmentation, serving to inhibit the formation of NBZs and shallow magma reservoirs [Head and Wilson, 1986]. For a range of common terrestrial magma volatile contents (<0.5 wt% H₂O, <0.35 wt% CO₂), magma ascending and erupting near or below the mean planetary radius (MPR) on Venus should not stall to produce shallow magma reservoirs. In this case, magma should ascend directly to the surface; such eruptions should be characterized by relatively high total volumes and effusion rates, comparable to those observed in terrestrial flood-basalt provinces (Figure 3) [Head and Wilson, 1992b].

Because atmospheric pressure changes considerably with elevation on Venus, the same range of volatile contents results in the production of NBZs and magma reservoirs at elevations well above MPR. For the same range of volatile contents at higher elevations (about 2 km above MPR), about half of the cases treated by Head and Wilson [1992b] result in direct ascent of magma to the surface and half in the production of NBZs. In general, NBZs and shallow magma reservoirs on Venus are predicted to appear as gas content increases and, because of the high atmospheric pressure, to be nominally shallower on Venus than on Earth. The shallowest depths for NBZs are about 1 km and depths increase slowly with increasing CO₂ content and rapidly with increasing H₂O content. For a fixed volatile content, NBZs become deeper with increasing elevation. Over the range of elevations (~1 to +4.5 km) treated by Head and Wilson [1992b], depths differ by a factor of 2–4, which is about the same factor as that induced by variations in CO₂. NBZ reservoirs can become deeper than reservoirs on Earth produced with similar volatile contents if common terrestrial volatile contents are exceeded. To a
first order, the characteristics and global distribution of volcanic landforms that are largely extrusive [Keddie and Head, 1994a, b] and structures that reflect intrusive activity [Grosfils and Head, 1995] support the idea that neutral buoyancy contributes to major aspects of volcano growth and development on Venus.

How do these different conditions influence the formation of LIPs? Populating the >80% of the surface of Venus comprised of volcanic plains are more than 1500 edifices or volcanic sources in excess of 20 km diameter [Head et al., 1992]. Over 150 of these are major shield volcanoes (Figure 5d) in excess of 100 km in diameter. The lack of a
hydrosphere and insignificant erosion on Venus mean that the early record of the volcano may be exposed and that different phases in its evolution can sometimes be more readily outlined than is commonly the case on Earth, particularly where flow lengths have decreased with time, leaving exposed sequential phases of volcano evolution [e.g., Keddie and Head, 1994a]. The heights of these shield volcanoes are considerably less than those on the Earth and Mars, typically less than about 2 km above the surrounding plains [Keddie and Head, 1994b]. Several factors help to account for these differences and illustrate how LIPs may be produced on Venus. One, the environment on Venus (surface temperature and pressure) favors larger primary magma reservoirs which will cause the wide dispersal of conduits that build edifices, resulting in broader, flatter structures. Two, models of shallow NBZ reservoir locations during edifice growth show that, for Earth, the center of the magma chamber remains at a constant depth below the growing summit of the edifice, thus keeping pace with the increasing elevation [Head and Wilson, 1992b]. In contrast, on Venus, because of the major gradient in atmospheric pressure with altitude, the chamber's center becomes deeper relative to the summit of the growing edifice. Although the chamber's elevation does rise with time, the rise rate is low. Therefore, magma reservoirs on Venus will remain in the pre-volcano substrate longer, and in many cases may not emerge into the edifice at all. In addition, the lower rate of vertical migration implies that, for a given magma supply rate, magma reservoirs would tend to stabilize, undergo greater lateral growth, and become larger on Venus than Earth. Thus, the proportion of the available magma going into production of the edifice relative to that intruded into the substrate is smaller on Venus than Earth. The resulting large reservoirs would encourage multiple and more widely dispersed source vents and large volumes for individual eruptions. All of these factors result in volcanic edifices that are low and broad, with reservoirs predomi-
nantly in the substrate, rather than the edifice. Because of the inhibition of volatile exsolution in the terrestrial sub-marine environment and its influence on volcanic landforms [e.g., Head et al., 1996b], these factors could also be important in the formation of LIPs on the seafloor and in the initial stages of Hawaiian-type edifice formation.

On Venus, over 150 large radiating lineament systems with a radius in excess of 100 km have been mapped [Grosfils and Head, 1994]. These structures are interpreted to be the surface manifestations of dike swarms radiating away from a central source, analogous to the giant radiating dike swarms on the Earth such as the Mackenzie dike swarm in Canada. On Venus, however, lack of erosion permits the surface equivalent of the deeply eroded Mackenzie-type swarms to be studied [e.g., Ernst et al., 1995]. Theoretical analysis and predictions of the characteristics of dikes emplaced in unbuffered (declining
driving pressure) and buffered (constant driving pressure) environments [Parfitt and Head, 1993] show that buffered driving pressure can readily account for the very wide (exceeding 100 m) and very long (in excess of 2000 km) dikes observed on the Earth and Venus in some mafic dike swarms. In addition, some dikes emplaced in buffered conditions will grow vertically and laterally until they reach the surface at some distance from the magma reservoir. The high driving pressure and large dike widths typical of these conditions mean that eruptions produced would be characterized by large volumes of lavas emplaced at very high rates, potentially producing flood basalts [Parfitt and Head, 1993] (Figure 5c).

An interesting class of features with associated large-scale effusive activity was discovered in Earth-based and Venera mission radar images. Named coronae, these features range in size from 60 to over 2000 km in diameter and are characterized by circular to elongate outlines and one or more discontinuous annuli comprising concentric compressional and/or extensional troughs and ridges [Pronin and Stefan, 1990]. Magellan provided data for a global census of coronae [Stofan et al., 1992; Head et al., 1992] that revealed over 300 of these features widely distributed over the surface [Magee Roberts and Head, 1993] and showed the details of their structure [Stofan et al., 1992; Squyres et al., 1992]. The prevailing interpretation is that these features represent the surface manifestation of hotspots or plumes [e.g., Stofan et al., 1991, 1992; Squyres et al., 1992; James et al., 1992] and that their diversity in structure and associated volcanism reflects differences in plume size and intensity and the thermal structure of the overlying lithosphere [e.g., Erickson and Arkani-Hamed, 1992]. Ascending mantle diapirs elevate and deform the lithosphere as they approach the base of the lithosphere. There, they flatten, spread laterally, and cool; on the surface the raised plateau subsides to produce a central depression and a deformed annulus and moat. One of the most impressive aspects of the documentation of these features is that they potentially represent hundreds of examples of hotspots whose history can be mapped because of the lack of erosion and incomplete coverage by later deposits.

A step in this direction was taken by Magee Roberts and Head [1993], who studied the temporal and spatial relationships of areally extensive volcanic flow fields associated with coronae. They showed that large-scale flow fields (average of $1.1 \times 10^5$ km$^2$, but up to $1.5 \times 10^6$ km$^2$; Figure 3) formed a significant stage in the evolution of at least 41% of all coronae and that the timing and scale of many coronae flow fields are consistent with the arrival and pressure-release melting of material in the head of a mantle plume. They also showed that those coronae with associated large-scale volcanic activity were preferentially located in areas of rifts (Figure 5c) and interpreted this to mean that the intersection of mantle upwellings with zones of extension is the most significant factor in the volume of melt produced and erupted from a corona.

Among the features observed on the volcanic plains of Venus are sinuous channels [Baker et al., 1992], many of which resemble sinuous rilles on the Moon. Lunar sinuous rilles, which range up to about 300 km in length, are thought to have formed from thermal erosion associated with high-effusion-rate eruptions [e.g., Hulme, 1982]. In some cases on Venus the deposits from the lavas that are proposed to have eroded the sinuous channels can be identified and distinguished from the surrounding plains. Many of the channels are much longer than 300 km and, indeed, one of them, Baltis Vallis (Figure 5b), is over 6800 km in length! These extreme lengths imply very high effusion rates and large-volume eruptions and may indicate that the lavas associated with these features were of an unusual composition, temperature, and/or viscosity (e.g., possibly komatiites [see Head et al., 1994] or carbonatites [Kargel et al., 1994]). Clearly, the eruption products forming these channels would qualify as LIPs.

Large lava flow fields approaching and exceeding the dimensions of many terrestrial LIPs are observed in numerous other settings on Venus. Mylitta Flectus is one such field that originates along a rift zone and covers about $3 \times 10^5$ km$^2$, extending down into Lavinia Planitia, a lowland basin. Maximum flow lengths range from 400 to 1000 km, flow widths from 30 to 100 km, and the total
volume of the flow field is of the order of $2 \times 10^4$ km$^3$\cite{MageeRobertsetal1992}. Magee and Head\cite{MageeHead1995} documented the morphology, morphometry, stratigraphy and distribution of the global population of large flow fields on Venus in excess of $5 \times 10^4$ km$^2$. The largest of the 208 such flow fields is $1.6 \times 10^6$ km$^2$ and the average area is $2.2 \times 10^5$ km$^2$ (Figure 3); collectively, the flow fields cover an area of $4.0 \times 10^7$ km$^2$, about 11% of the plains regions of Venus. The most common source vents for the large flow fields are coronae, large volcanic shields, and fissures and fractures within rifts and fracture belts. Most flow fields are associated with zones of extension, such as major rift zones and fracture belts, and the emplacement of the flow fields tended to postdate the onset of extension. In reference to terrestrial flood basalts and the discussion about the relative importance of large-scale mantle upwelling (e.g., plume heads) versus lithospheric extension causing enhanced decompressional melting, the Venus data support the idea that lithospheric extension and thinning accompany the formation of the majority of flood-basalt lavas there. In addition, examination of a 6800-km-long rift zone interpreted to have originated from passive rifting in response to stresses linked to adjacent downwelling shows that extension occurred generally prior to the eruption of large-scale volcanic flow fields, comparable to some terrestrial flood basalts\cite{MageeRobertsetal1992,MageeHead1995}. This is in contrast to the Columbia River and Deccan Basalt Groups, where evidence has been presented that eruption of the main tholeiitic phase preceded significant extension and crustal thinning\cite{Hooper1990}.

Many rift zones on Venus are associated with broad rises resembling the Tharsis and Elysium regions on Mars. The Beta, Atla, and Western Eistla regions are each up to 2000-3000 km in diameter and rise up to several kilometers above the surrounding plains (Figure 5a). They are characterized by rift systems which cross (Western Eistla) or radiate away from (Beta and Atla) the central high, and large shield volcanoes are located on the summit and flanks of the rise. Positive gravity anomalies are consistent with mantle upwelling\cite{Sensketal1992}. These rises appear to represent a scale of mantle upwelling much larger than that related to individual volcanoes and coronae\cite{Headetal1992}, although several coronae reach tremendous dimensions, e.g., Heng-O, 1060 km, and Artemis, ~2500 km. The global distribution of volcanic landforms revealed by the Magellan mission showed a concentration of volcanic edifices and sources in areas comprising about $9.2 \times 10^7$ km$^2$, or about 20% of the surface of Venus\cite{Headetal1992,Crumpleretal1993,1996b}. This area of regionally abundant volcanic sources also corresponds to the location of three major rifted rises, Beta Regio, Atla Regio, and Themis Regio (thus the term BAT region). This concentration suggests several scales of upwelling and instabilities (relatively small for the individual volcanic source regions, a thousand kilometers for the broad rises, and perhaps many thousands of kilometers for the BAT region). The BAT region covers a comparable planetary surface area percentage (20%) to that of the Tharsis rise on Mars. The ages of its individual components (riifting, rises, volcanoes) largely postdate the earliest plains emplacement\cite{Crumpleretal1994,BasilevskyHead1995a,b}, and thus it may represent mantle convection patterns linked to the aftermath of the collapse of a negatively buoyant, depleted mantle layer remaining from the extraction of the basaltic crust\cite{Head1995}.

Preliminary analysis of the global stratigraphy of Venus suggests that the dominant geologic processes and styles of volcanism have changed over time\cite{BasilevskyHead1995a,b,1996}. The oldest terrain exposed is known as tessera (tile in Greek, for the similarity of the terrain texture to parquet floor tiles). This terrain is high-standing and very complexly deformed, somewhat continent-like, and comprises about 8% of the surface of Venus\cite{IvanovHead1996}. Tessera is embayed by two major plains units and is thus probably much more widespread in the subsurface than its present outcrop would suggest. The oldest of the two major plains units (ridged plains) covers most of the surface of Venus. Most ridged plains are homogeneous and flow-unit boundaries generally are not traceable; however, ridged plains are characterized by numerous sinuous channels, suggesting large-volume, high-effusion-rate eruptions. The stratigraphically younger regional plains unit is characterized by smoother lobate flows (such as those at Mylitta Fluctus) usually emanating from discrete sources. Although individual flows are volumetrically significant and often akin to flood basalts, they do not have the distinctive sinuous channels of the ridged plains and thus appear to have a different mode of emplacement. Large volcanic edifices representing individual volcanic sources, and the emplacement of hundreds of flows from subjacent localized magma reservoirs are the most recent features and these are superposed on most other types of plains units\cite{Crumpleretal1997}. Thus, following tessera formation, two units that could be interpreted as LIP-related were formed: the ridged plains, where large-scale, sinuous channel-type emplacement occurred, and the smooth/lobate plains, where flood-basalt-like provinces were produced, often in conjunction with rift zones. This major change over time (together with substantial changes observed in the thermal
evolution of Mars and the Moon) suggests that the style, number, and size of LIPs may vary over the long-term geologic record of the Earth.

What causes these changes on Venus? The size-frequency distribution of exposed impact craters shows that the average surface age (crater retention age) is about 300–500 Ma, much more similar to that of the Earth than the smaller terrestrial planets (Figure 1). Even more surprisingly, the areal distribution of craters cannot be distinguished from a completely spatially random population [Schaber et al., 1992; Phillips et al., 1992; Strom et al., 1994] and nearly all of the craters have not been modified by post-emplacement volcanism. On the basis of these results, it was hypothesized that Venus underwent a global tectonic and volcanic resurfacing event about 300–500 m.y. ago that eradicated the previous cratering record. Subsequent to that event (thought to have lasted about 10^7–10^8 years), volcanism was relatively minor in volume and areal distribution (on the basis of the small number of craters modified by volcanic activity) [Schaber et al., 1992; Strom et al., 1994]. Many mechanisms have been proposed to explain this hypothesized event [e.g., see review of Solomon, 1993], including episodic plate tectonics [e.g., Turcotte, 1993]. Among these hypotheses [e.g., Schaber et al., 1992; Parmentier and Hess, 1992; Head et al., 1994] are mechanisms that call for near-global volcanic resurfacing, in effect a planet-wide LIP! In the scenario proposed by Parmentier and Hess [1992], vertical crustal accretion leads to formation of a thick, melt-depleted mantle layer that evolves chemically and thermally over geologic time; the depleted mantle layer ultimately becomes negatively buoyant and founders. This event is predicted to occur over a geologically short time [Parmentier and Hess, 1992] and the founding, downwelling depleted mantle layer is hypothesized to have deformed much of the crust into tessera terrain, while the complementary upwelling fertile mantle underwent massive pressure-release melting to produce voluminous sinuous-channel-related flood basalts over a relatively short time (≤10^8 yrs) [Head et al., 1994]. Subsequently, volcanism waned but was locally significant in rifts (where local flood basalt units were emplaced, such as Mlyitta Fluctus) and near hotspots, where magma reservoirs evolved to produce volcanic edifices (e.g., Sapas Mons).

Three important observations can be made relative to the study of LIPs on Earth. One, the hypothesized depleted-mantle-layer overturn event on Venus could be the equivalent of a planet-wide LIP. Approximately 80% of the planet (3.68 × 10^8 km^2; Figure 3) may have been resurfaced over a very short time during the emplacement of the ridged plains [Basilevsky and Head, 1995a, b]. Although volumes are not well known, they may be of the order of 9.2 × 10^4 km^3 [Head et al., 1992]. Effusion rates are likely to have been very high for many plains units emplaced through sinuous channels, but estimates are difficult to make. Integrated fluxes were very high, however. If the ridged plains were emplaced in about 10^8 years, the average flux would be 5–7 km³/a (Figure 3), which is about three orders of magnitude more than the peak lunar mare flux, approximately comparable to the typical extrusive component of the Earth at present (intraplate and plate boundary) and more than five times greater than the flux typical of the last tens of millions of years for Venus. Two, the great contrast in magma flux between the resurfacing event and subsequent activity shows that major changes can take place in the geologic history of a planet; different types of LIPs can occur in relatively rapid succession, and periods can occur when virtually none are emplaced. Three, if either the hypothesis concerning the buildup and collapse of the depleted mantle layer or the episodic-plate-tectonic hypothesis is correct, this means that large-scale planetary heat loss can be cataclysmic and episodic, a phenomenon not considered in monotonic thermal evolution models.

One implication of the tectonic and volcanic record of Venus is that the crust forms and evolves [Head, 1990b] primarily in a vertical sense rather than in a lateral sense, as is the case in terrestrial oceanic plate spreading, although hypotheses for episodic plate tectonics on Venus have been proposed [e.g., Turcotte, 1993]. This concept of vertical crustal accretion has important implications for the production of LIPs on Venus and the general evolution of secondary crust over geologic time [Head et al., 1994], as discussed below.

Other Planetary Bodies

Outside the orbits of the terrestrial planets lie the asteroid belt and the outer gas giant planets and their satellites. Some meteorites and asteroids show evidence for differentiation and basaltic volcanism [e.g., Taylor et al., 1993], phases of which may have been volumetrically significant [e.g., Wilson and Keil, 1996]. Outer planet satellites are predominantly low-density bodies composed primarily of water and related ices [e.g., Burns and Matthews, 1986]. One exception is the innermost of the Galilean satellites of Jupiter, Io, which is approximately the same size and density as the Earth's moon. In one of the most spectacular predictions [Peale et al., 1979] and discoveries [Morabito et al., 1979] of planetary exploration, images returned by Voyager showed numerous active volcanic eruptions on Io [Smith et al., 1979].
Through a combination of pyroclastic eruptions and lava flows, Io appears to be resurfaced at the phenomenally high rate of $10^4$ to 1 cm/yr [Nash et al., 1986]. Further exploration by the Galileo mission will provide evidence for the nature of changes on Io in the last 17 years and the relation of these resurfacing rates and styles to LIPs.

**SUMMARY, RELEVANCE TO TERRESTRIAL LIPS, AND OUTSTANDING QUESTIONS**

**Environments, Associations, Settings of Formation and Style of Emplacement**

The planetary record provides a perspective on the three main categories of terrestrial LIPs: oceanic plateaus, continental flood basalts, and volcanic passive margins. Multiple analogs to continental flood basalts (e.g., Moon) and to oceanic plateaus (e.g., Mars, Venus) exist on the terrestrial planets; the rift-related LIPs on Venus provide important information about sequence and timing of emplacement in relation to volcanic passive margins. The role of large-volume, long-distance, lateral dike emplacement of flood basalts is illustrated by coronae on Venus; these provide probable analogs to the now-eroded giant radiating dike swarms of Earth and associated, but largely eroded, flood basalts. The great range of scales of upwellings on Venus and Mars, the influence of crustal thickness and composition, and lithospheric thickness variations in space and time also are significant for studies of terrestrial environments. Venus shows the potential significance of vertical crustal accretion and the influence of the complementary depleted mantle layer on further petrogenetic evolution, as well as the possibility of episodic, cataclysmic planetary-scale resurfacing. Theoretical analysis of environments on Venus also illustrates that differences in thermal structure can cause fundamental differences in the volume of melt produced in buoyant upwellings in both rise [e.g., Sotin et al., 1989] and plume [e.g., Erickson and Arkani-Hamed, 1992] environments. It has been proposed that many terrestrial LIPs are analogous to the lunar maria and resulted from impact-related pressure-release melting [e.g., Alt et al., 1988]. The lunar geologic record shows that this is unlikely at the present time because of the generally small size of impact projectiles, but that it may have been more significant in past Earth history when larger impacts occurred, especially in oceanic settings.

**Influence of Surface Environment**

The example of Venus suggests that the external environment into which magma is extruded can have a major influence on the occurrence, depth, and size of magma reservoirs and thus on the possibility of flood basalts [Head and Wilson, 1992b]. This consideration implies that intraplate submarine reservoirs and extrusions (more Venus-like) may be different from those in subaerial environments on Earth, as indeed might many LIPs formed earlier in Earth's history, when atmospheric pressure may have been higher.

**Controls on Mode and Location**

The lunar record shows how a low-density crustal layer analogous to continental crust on Earth can act as a filter to plumes and associated volcanic activity, obscuring their surface manifestation and even precluding the construction of shallow magma reservoirs and large shield volcanoes. Variations in lithospheric thickness on the terrestrial planets illustrate how thermal structure can influence the occurrence and mode of large-volume eruptions and how changes in lithospheric thickness with time can alter the style and abundance of flood basalts.

**Relation to Internal Structure**

The wide range of plume-like features with associated large extrusive components suggests that plume sources could possibly extend from the upper mantle to the core-mantle boundary. The extremely large and long-lasting provinces on Mars (Tharsis) and Venus (BAT) strongly suggest that even larger instabilities occur than those commonly associated with individual terrestrial plumes. Better knowledge of variations in the volcanic flux associated with plume-like features on Venus will help us to understand terrestrial plume and mantle structure [e.g., Bercovici and Mahoney, 1994]. Another potentially important perspective comes from the vertical crustal accretion hypothesis for Venus; the formation and accumulation of a complementary melt-depleted mantle layer can significantly alter the nature of further melt production and can also influence crustal buoyancy and stability. The lunar highland crust density barrier also illustrates that in some cases basalts may localize at the base of the crust, as is believed to occur for terrestrial flood basalts magmas. Venus demonstrates that large-scale melting in the mantle (the hypothesized large-scale mantle overturn) may have had very long-lasting effects on mantle convection patterns and volcanism [e.g., Crumpler et al., 1993].

**Implications for Plume Structure**

Although the compositional, thermal and mechanical structure of the crust and lithosphere of the planets is not
the same as on the Earth today, the planetary record can provide a frame of reference for questions such as plume structure (plume heads and tails), plume incubation versus plume impact [e.g., Kent et al., 1992], internal mantle structure [e.g., Bercovici and Mahoney, 1994], and plume duration (e.g., on Venus and Mars).

Large Igneous Province Substructure

Eroded, giant radiating dike swarms on Earth (e.g., the Mackenzie dike swarm) and their uneroded counterparts on Venus [e.g., Ernst et al., 1995] show that flood basalts need not occur only above a plume head; dike thicknesses, lengths, and flow rates are such that flood basalts can occur several thousand kilometers away from a plume, through lateral transport of magma in dikes and its eruption due to buffered conditions in the magma reservoir [Parfitt and Head, 1993].

Areas and Volumes

Planetary LIPs have a wide range of areas and volumes (Figure 3), showing that terrestrial LIPs are not unique in this respect. In addition, volumes range up to that of Tharsis on Mars, and scales on Venus exceed the present surface area of the oceanic crust. These examples suggest the possibility of larger terrestrial LIPs than presently recognized (e.g., terrestrial superplumes [Larson, 1991a, b]). The volumes of the larger planetary LIPs and the stationary lithosphere of most terrestrial planets suggest that voluminous partial melting of mantle has occurred (e.g., see figure 10 of Coffin and Eldholm [1994]) and that residual, depleted mantle layers must play an important role in the continued evolution of planetary upper mantles.

Duration and Rates of Emplacement

Planetary LIPs are seen in which volumes were very high and eruption durations were both short (the large outflows and sinuous rilles on the Moon) and long (the shield volcanoes on Mars). In addition, the Tharsis rise on Mars shows that mantle melting anomalies can last billions of years, producing prodigious LIPs, and the Venus global resurfacing model suggests that large-scale mantle overturn may provide short-term ($\leq 10^5$ yr) pulses of global-scale igneous provinces. On the basis of the planetary perspective, high eruption rates may be due to a variety of conditions, including trapping of melt at density barriers and subsequent overpressurization of reservoirs [Head and Wilson, 1992a], buffered conditions in magma reservoirs [Parfitt and Head, 1993], higher temperatures and degrees of melting in the earlier history of planets, periods of possible cataclysmic resurfacing (e.g., on Venus [Head et al., 1994]), and periods of anomalous rates of mantle convection. The cataclysmic resurfacing hypothesis for Venus also illustrates the possibility that thermal evolution may not be steady-state and monotonic but rather episodic [see also Condie, 1995].

Petrogenetic Evolution

Do extraterrestrial LIPs contain differentiates of basalt, and if so, where do these occur in the sequence? In situ geochemical analyses on Venus suggest that tholeiites and possibly more alkaline basalts [Sukov et al., 1987] form the vast volcanic plains. The extensive sinuous channels have been interpreted as evidence for possible komatiites [Head et al., 1994] and carbonatites [Kargel et al., 1994]. Steep-sided domes [Pavri et al., 1992] and large deposits of viscous-appearing deposits [Moore et al., 1992] have been observed. Unfortunately, widespread and detailed chemical analyses have not yet been made on the planets.

Influence on the Atmosphere and Environment

Voluminous and prolonged volcanic outpourings can make important contributions to the atmosphere of planets throughout their evolution, as on Mars [e.g., Greeley, 1987]. If large volumes of flood basalts are extruded over very short periods, outgassed volatiles, heat flux, and voluminous particulate matter can influence short-term chemistry and circulation of the atmosphere and long-term climate evolution. Potentially the most dramatic example of this is the widespread volcanic resurfacing hypothesized for Venus. For example, Bullock and Grinspoon [1996] showed that an increased flux of volcanism such as that interpreted to be associated with the proposed global resurfacing would precipitate a climatic catastrophe leading to much higher temperatures and pressures.

Relation to Geologic History

A most important perspective from the planets is that the characteristics and rates of geologic processes as a function of time and thermal evolution have experienced large-scale changes. The geological processes dominating the geologic record over the last several hundred million years on the planets (i.e., the temporal equivalent of the Phanerozoic on the Earth) are not the same as those operating in the earlier history of Mars, Venus, Mercury, and the Moon. Thus, we should anticipate potentially major changes in the style of volcanic extrusion as a function of
geologic time on Earth. It is clear from the planetary record that coincident with the general thermal evolution of the planet, changes can occur in the mantle convection planform, the development and scale of mantle instabilities, and the conditions of melting in the Earth's crust. Much of the evidence from the planets and from thermal evolution models suggests a more important role for LIPs in the earlier history of the Earth.

**Origin of LIPs: A Planetary Perspective**

The planetary perspective provides many examples of LIPs in a diverse range of geological environments. This underlines the fact that there is no single origin for LIPs, but that, taken together, they can help to understand the nature and significance of large-scale melting in the shallow interiors of planets [e.g., Coffin and Eldholm, 1994]. For example, the broad rifted rises of Venus and Mars and their associated LIPs serve as potential analogs for early continental breakup and the early stages of crustal spreading on Earth. Continued analysis of data from the planetary record will help to provide perspective on terrestrial LIP dimensions, durations, and rates of emplacement, as well as mantle and crustal structure and processes, relationship to tectonism, environmental effects and petrological and geochemical characteristics and evolution.

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