Surfaces and Interiors of the Terrestrial Planets

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A schematic cross-section through Earth reveals its immense iron core.

IN THE LAST 35 years, two parallel revolutions in the Earth sciences have radically altered our perceptions of planets and how they work. The first was the development of the theory of terrestrial plate tectonism. During the 1960s we came to realize that Earth's geologic expressions are not an assortment of isolated puzzles to be solved individually, but rather a record of the movement and interaction of a small number of large lithospheric plates operating in an integrated, global manner. The second revolution resulted from the unfolding view, thanks to spacecraft exploration, of planetary surface features. Almost overnight, it seems, the Moon and planets changed from astronomical objects to geologic ones. In doing so, they began

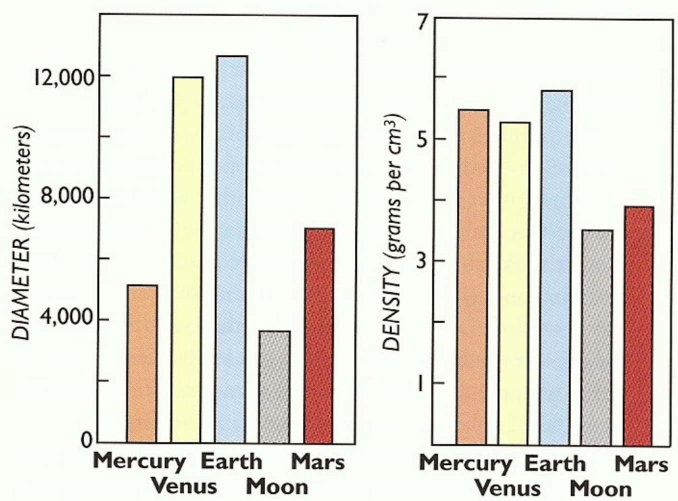


Figure 1. Mercury, Venus, Earth, Mars, and the Moon are crudely comparable in size, and all five have relatively high densities. Collectively known as terrestrial planetary bodies, their silicate-dominated compositions are distinct from those of the outer planets and their satellites, which are rich in ices and other volatile elements.

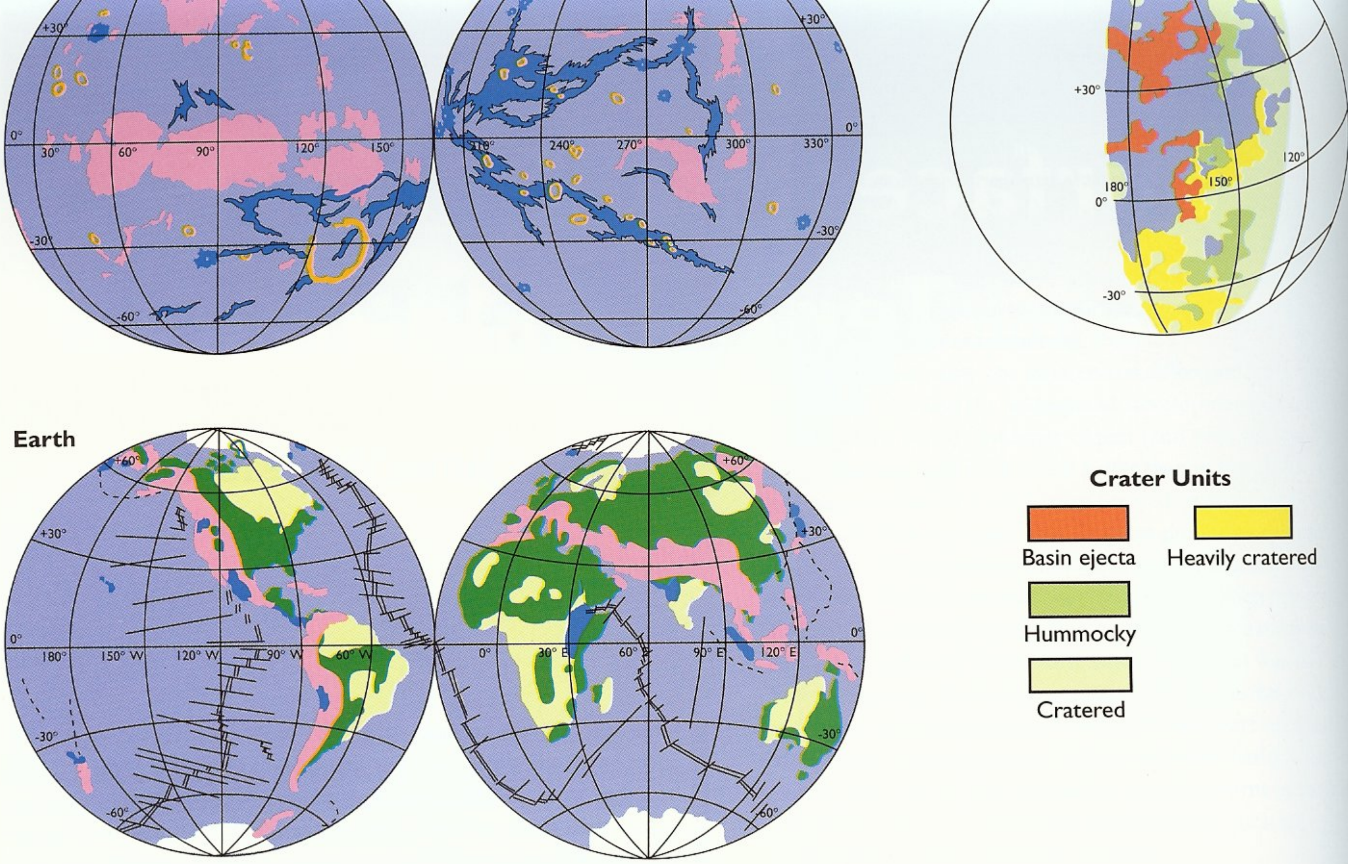


Figure 2. **These generalized geologic maps of terrestrial planetary bodies bring together data from numerous spacecraft investigations and a variety of other sources. Several geologic surface types are common to all five worlds, including cratered terrain and volcanic plains. Others are limited to one or a few of them, such as polar deposits (Earth, Mars), tesserae and coronae (Venus), and folded mountains (Earth, Venus). In general, the surfaces of the Moon and Mercury are more primitive and less complex than those of Earth, Venus, and Mars. Of the five worlds, only the Earth and Venus are believed to have undergone extensive surface modification in the last 2 billion years. One puzzling feature common to Earth, Mars, and the Moon is a dichotomy of terrains that results in a hemispherical asymmetry. On Earth, the dichotomy is between the continents and the volcanic plains of its ocean basins; on the Moon it is between ancient, heavily cratered highlands and mare plains (which are concentrated on the lunar near side); on Mars it is represented by northern lowlands and southern cratered uplands. The origins of these hemispherical asymmetries are not fully understood.**

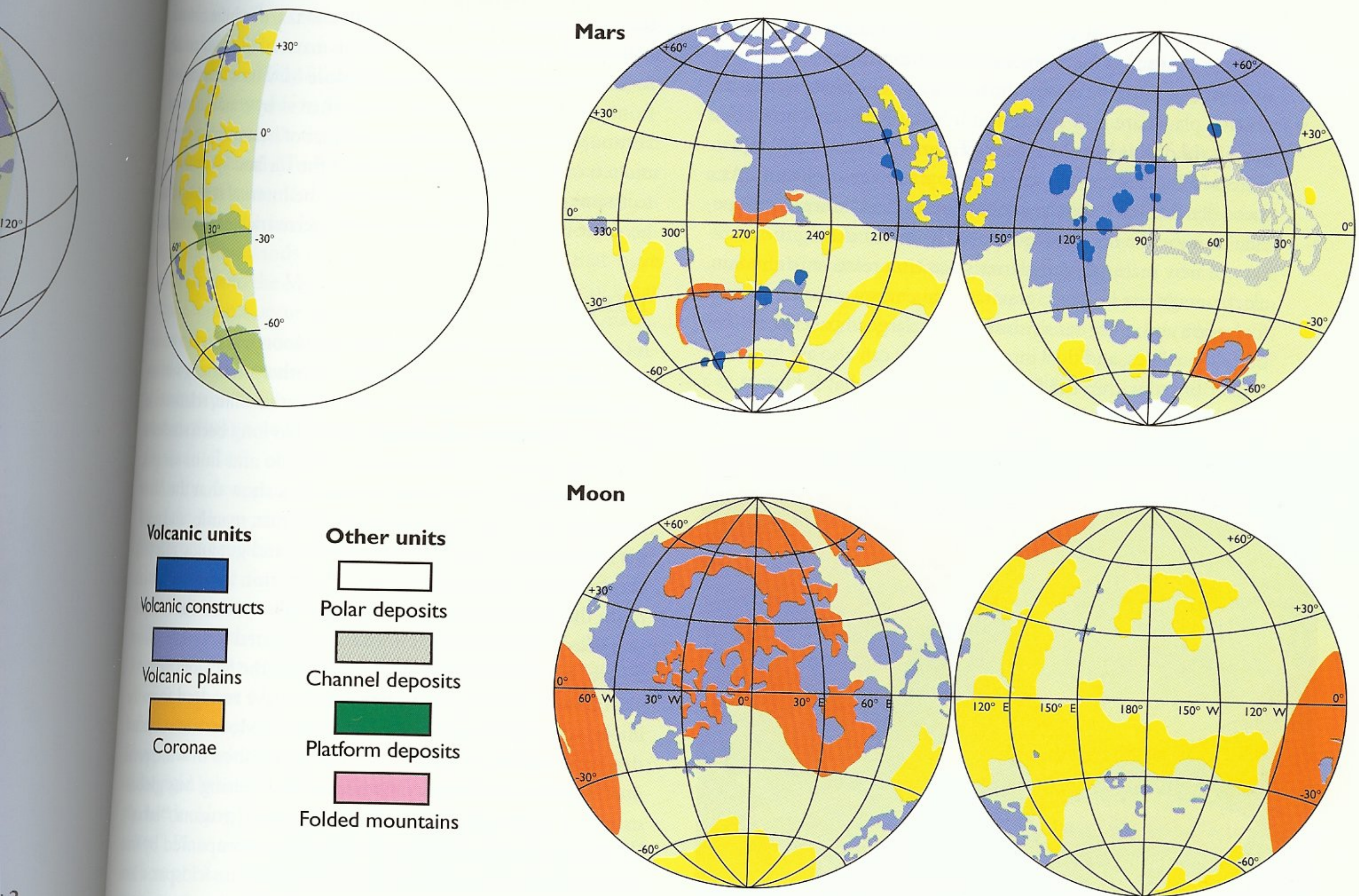
to provide a framework within which Earth could be viewed not as a single data point, but as one of a family of planets.

The point is that, prior to the 1960s, geologic thought focused on Earth and in particular on specific areas of its surface. In a rather passive sense, this mindset was analogous to the pre-Copernican, Earth-centered view of the solar system. Now we have abandoned such geologic chauvinism, working and thinking instead in terms of a new solar system in which Earth's history is inextricably linked to those of the Moon and the other terrestrial planets. During the 1990s, Earth scientists began to apply their understanding of recent geologic history to the first 4 billion or so years of our planet's existence. Simultaneously, others consolidated their understanding of the formative years of planetary evolution (the first 1½ billion years) and began applying it to Earth. The study of Venus will provide a driving force for this convergence. It is the most Earth-like of the terrestrial planets in terms of gross physical properties (*Figure 1*), and we must learn to what extent it does — or does not — mimic Earth in detail.

THE INGREDIENTS FOR PLANETARY EVOLUTION

Three fundamental processes — impact cratering, volcanism, and tectonism (crustal movement) — give planetary surfaces most of their characteristic features (*Figure 2*). In addition, as later sections will show, each terrestrial planet expresses these processes with a unique signature. Our studies of the surfaces of these worlds reveal an amazing diversity of characteristics, a few emerging themes, and a host of basic new questions.

However, we cannot address questions of the planets' evolution only by studying their surfaces, any more than we can tell the life story of a group of human beings just by looking at them. Like humans, planets are complex systems in which most of the driving forces and regulating processes are hidden below the surface. To study these forces and processes, indirect measurements are required. Indeed, the surface features must be studied in detail to infer the nature of the interior and how it might have changed with time.



Once we know the answers to basic questions — whether a planet's interior is homogeneous or layered, whether heat in the interior has been high enough to cause volcanic activity, and so on — we can ask more sophisticated questions, such as: How do planets differ in their basic internal structure, and how does this determine their evolution? What role do a planet's size, and its position in the solar nebula during its formation, play in its further development? Questions like these are basic tools of comparative planetology, which views the solar system as a single "experiment" in which differences among planets can be explained by variations in initial composition, distance from the Sun, and so on.

The most basic process that shapes the early evolution of a planet is ***differentiation***, the segregation of a relatively homogeneous interior into layers or pockets of different composition (*Figures 3,4*). There is ample evidence that each of the terrestrial planets, as well as Earth's Moon, has undergone this segregation, producing a crust, mantle, and core. Differentiation can occur rather rapidly and be rather catastrophic. The process of core formation, in which denser, iron-rich material sinks to the deepest interior, is thought to occur in the first few percent of a planet's history, when the interior is hottest. The amount of gravitational potential energy released by the process of core formation is incredibly high, enough to melt the entire surface of a body the size of Earth.

Volcanism and magmatic activity are essentially a second phase of differentiation that takes place over longer periods. Heat in the interior, caused by the decay of radioactive elements, causes partial melting of the mantle. The resulting magmas, which are hotter and less dense than their surroundings, rise toward the surface to form crustal intrusions and extrusions such as lava flows and volcanoes. The amounts and rates of lava production have varied from planet to planet (and over time on each individual planet). However, volcanism's influence on the surfaces of all five worlds has been widespread. For example, the dark lunar maria first appeared about 4 billion years ago and were almost completely emplaced about 1½ billion years later. Volcanic activity on Mars and the formation of its great shield volcanoes continued well into the most recent half of the solar-system history, and of course volcanic activity on Earth (and possibly Venus) occurs frequently even now.

In the early history of the solar system, impacts were another cause of differentiation. A meteoritic projectile strikes a planetary surface at velocities measured in kilometers per second. The kinetic energy concentrated at that point can easily equal the total annual heat flow of Earth! Consequently, the cataclysmic bombardment associated with the final stages of planetary formation was energetic enough to melt the outer parts of the Moon and terrestrial planets. In these global magma oceans, lighter mineral species floated to the top and formed the planet's primary crusts. Later internal heating and melting then produced materials for a secondary crust, and reworking of these earlier crusts has yielded tertiary crusts (*Figure 5*).

It is obvious that heat, which drives differentiation and subsequent volcanism, is the single most significant variable in shaping a planet's evolution. Among the most basic questions we can ask about a planet are how much heat it had initially, and how it has gotten rid of this heat over time (*Figure 6*). These simple questions are the keys to understanding planetary evolution. At a more detailed level, we can ask how much heat derives from specific sources: the energy acquired at the outset during the accretion of new material, the planet's position relative to the Sun, electromagnetic heating, gravitational energy released by core formation and other density instabilities, large impacts, the decay of radionuclides, and tidal interactions. We can also ask how heat is distributed within the interior over time, what is known as a planet's thermal history. We can estimate the rates of change of temperature as a function of depth and use this information to help predict the physical state of material (liquid, partially molten, or solid). And we can explore how heat is transferred from one part of the interior to another — by conduction, convection, or advection, the direct transfer of heat by movement of molten material from the interior to the surface. We will return to these questions after considering the internal structure and geologic history of the Moon and the terrestrial planets in detail.

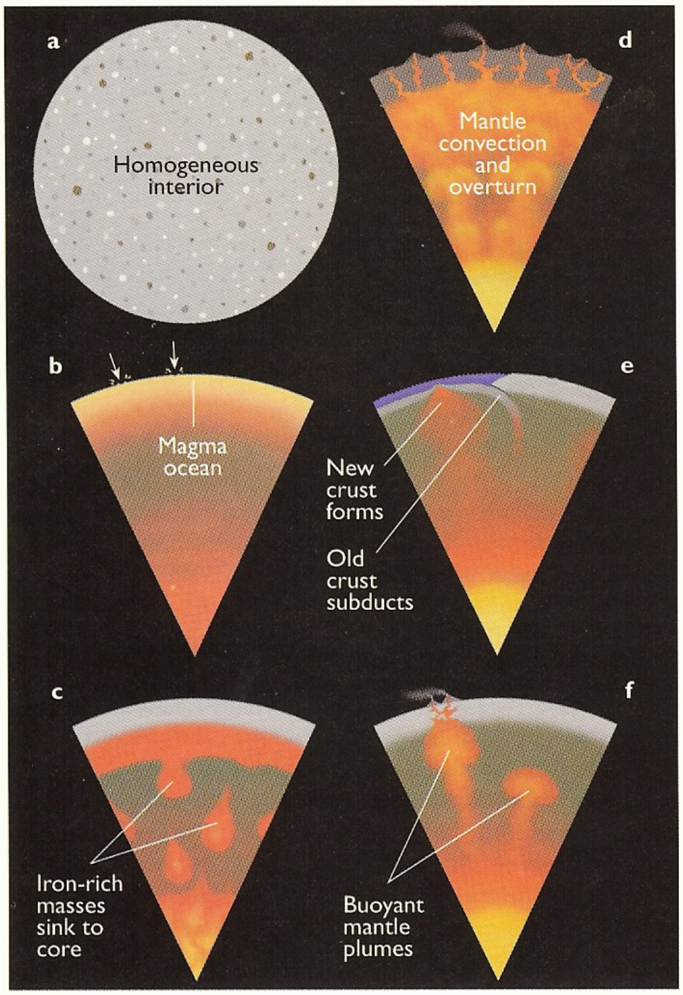


Figure 3. **Differentiation is the process by which a body's initially homogeneous primitive materials *(a)* become segregated in planetary interiors. A number of different mechanisms contribute to differentiation. For example, the kinetic energy delivered by impact bombardment *(b)* can cause widespread — even global — melting of near-surface layers. Denser material collects and sinks to form a core *(c)*, and the heat released from this event can trigger further differentiation. Deep-seated compositional and thermal instabilities *(d)* can cause materials to become less dense than their surroundings, resulting in rising plumes that undergo mixing and further differentiation. On Earth's ocean floors, differentiation also occurs as slabs of oceanic crust *(e)*, mixed with water and sediment, are forced down into the crust and remelted. On Venus *(f)*, vertical crustal accretion may continue until density instabilities cause the crust to founder, allowing planetwide resurfacing by volcanism.**

THE MOON

As our closest neighbor in space, the Moon was first to occupy the attention of geologists studying other planetary surfaces (see Chapter 10). The Moon is the largest satellite, relative to its parent body, in the solar system, and it has long been studied as if it were a planet in its own right. Apollo and Luna samples, remote sensing, and surface seismic data show that the Moon has been internally differentiated into a crust, mantle, and possibly a small core (*Figure 7*). Seismic data and geologic mapping show that the lunar maria are relatively thin (a few km maximum thickness) and perched on a globally continuous feldspar-rich crust. This crust is thinner on the central near side, about 55 km, but may be up to 100 km thick on the far side.

Following its formation, believed to have resulted from the impact of a Mars-sized body with Earth, the Moon was subjected to a period of heavy bombardment that resulted in widespread melting. Opinions differ on whether the melting was globally extensive (a magma "ocean") or regional (magma "lakes"). Whatever the exact details, melting was accompanied by fractional crystallization and separation of plagioclase feldspar, a low-density silicate mineral. This floated to the surface to create the primary crust, which survives today as the lunar highlands (*Figure 8*) and the residual upper-mantle layers. The residual layers below the crust were denser than the underlying mantle, which probably caused them to sink toward the interior and, perhaps, to form a core. During this latter period, partial melting of the mantle created magmas that flooded the lunar surface and

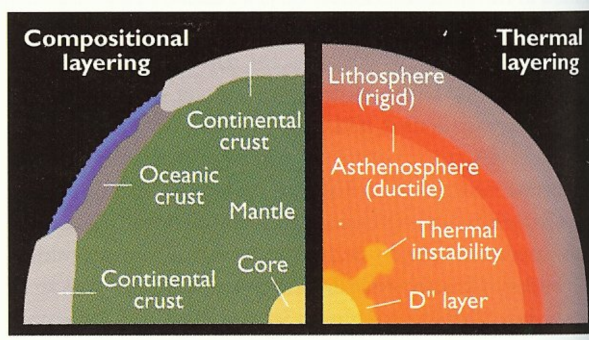


Figure 4. **Planetary bodies are commonly divided into compositionally discrete layers (left half), such as a dense, iron-dominated core, a mantle rich in iron- and magnesium-silicates, and a crust that is of primary, secondary, or tertiary origin (or a combination of these). Thermal layering (right half) is caused by changes in temperature that affect the properties of planetary materials. For example, within Earth's crust the boundary between the rigid lithosphere and the more plastic aesthenosphere occurs at a depth of about 100 km. Deeper down, the D" layer at the core-mantle boundary may trigger rising mantle plumes that produce surface hotspots.**

formed a secondary crust. By about 2½ to 3 billion years ago, basaltic lavas had covered approximately 17 percent of the lunar surface, preferentially filling the interiors of the giant impact basins on the near side to form the lunar maria (*Figure 9*).

Tectonic activity on the Moon stands in stark contrast to that of our own planet. The Moon lacks any evidence of the major lateral crustal movement so typical of Earth. Instead, lunar tectonism is characterized by downward movement, due to loading by the thousands of cubic kilometers of lava that have poured out onto the Moon's surface. The limited array of lunar tectonic features occurs predominantly in and near the maria. Linear rilles and graben, formed by crustal extension, were followed by sinuous (wrinkle) ridges formed by contraction.

Instruments left by the Apollo astronauts found that the amount of heat now flowing from the lunar interior is far less than that of Earth, consistent with a body that is losing heat by conduction. It appears that at present the outer 800 to 1,000 km of the Moon acts as a relatively rigid shell, or *lithosphere*. However, the presence of the highlands crust, mare basalts, and related tectonic features show that earlier in its history the Moon had a hotter interior. How was this heat lost?

The Moon's thermal evolution was shaped most of all by the formation of the globally continuous primary crust (*Figure 10*). Because the Moon has a large surface area relative to its volume, cooling by conduction was very efficient and the lithosphere cooled and thickened rapidly. In contrast to Earth, whose rigid lithosphere became divided into a number of moving, interlocking plates, the Moon quickly became a "one-plate planet." Plate tectonism never developed, and the Moon has lost heat primarily by conduction ever since. The small fraction of the surface covered by the volcanic maria indicates that advective cooling played a minor role. Evidence from tectonic features suggests that the Moon underwent a change from a net global expansion prior to about 3.6 billion years ago to a net contraction that continues today.

Gravity data show that there are large concentrations of mass, or *mascons*, associated with the youngest mare-filled basins. The spacing and type of tectonic features around the mare margins provide evidence that the surface was flexing and subsiding due to the load of the lavas on the lithosphere. The mascons plausibly represent the last outpourings of mare lavas, emplaced on a lithosphere that was so thick that it was able to support this load. The lack of a significant dipole magnetic field at present, combined with evidence for a fossil magnetic field in some lunar samples, is consistent with this internal thermal evolution.

Virtually no major internally generated geologic activity has manifested itself on the lunar surface for the last 2½ billion years. The Moon preserves a record of the impact bombardment and volcanism that dominated the early solar system, and thus it serves as a Rosetta stone for interpreting the surface geology of other terrestrial planets.

MERCURY

Mercury is one of the most poorly studied and enigmatic planets (see Chapter 7). Clues to its unusual nature come from the fact that it is about one-third the diameter, but about the same density, as Earth (*Figure 1*). These characteristics offer an opportunity to study the influence of size and internal structure on a planet's geologic history and thermal evolution. Mercury also raises the question of whether a planet's initial starting conditions govern its evolution.

The Mariner 10 spacecraft returned images of about 35 percent of the planet's surface and at first glance revealed a lunarlike terrain (*Figure 11*). However, Mercury differs from the Moon in several important respects. Large areas of relatively ancient intercrater plains may indicate that early volcanism (during the period of heavy cratering) was more extensive on Mercury than on the Moon. Large, extended scarps on Mercury (*Figure 12*)

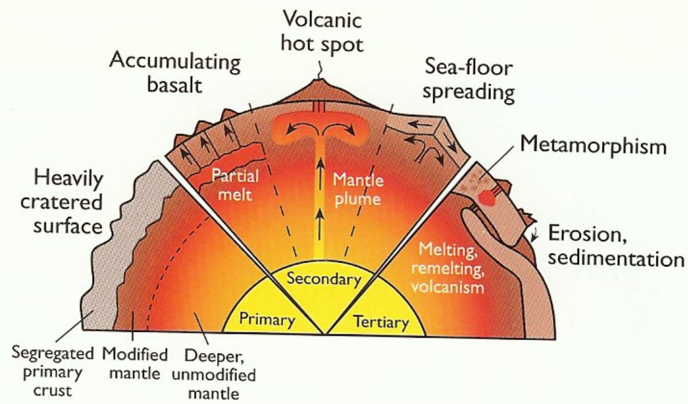


Figure 5. Various mechanisms lead to the formation of a planet's primary, secondary, and tertiary crusts. Primary crust, created early in planetary history, is still preserved in places like the lunar highlands. Partial melting of the mantle and volcanic activity lead to the formation of secondary crusts, largely of basaltic composition. Tertiary crust results from the recycling of primary and secondary crust, as is typified by Earth's continents and (perhaps) the tessera terrain of Venus.

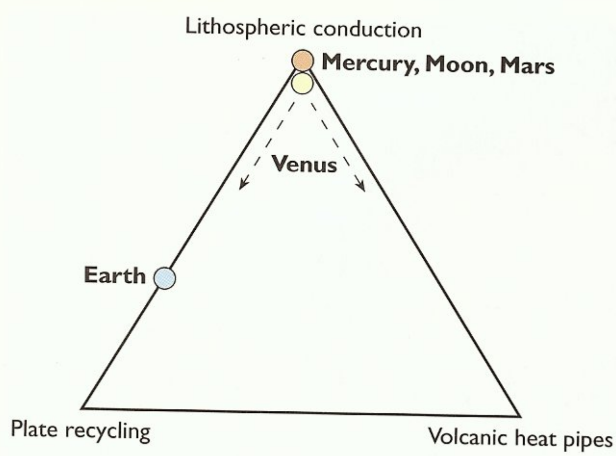


Figure 6. Methods of heat transfer within the terrestrial planetary bodies. At present, the Moon, Mercury and Mars lose their heat almost completely by the passive conduction through the lithosphere. Earth loses most of its heat through formation of the oceanic lithosphere at mid-ocean ridges, and through the spreading and cooling of this lithosphere. On Venus, conduction is the primary heat-loss mechanism today, though episodes of plate tectonism (left arrow) or widespread volcanic resurfacing (right arrow) may have occurred in the past. Advection, the direct transfer of heat by molten material to the surface through volcanic conduits, does not appear to be a significant process on any of these bodies. However, it almost completely dominates heat loss on the Jovian satellite Io.