Chapter 8
Endogenic Processes

Ernst Hauber, Daniel Mège, Thomas Platz, and Petr Brož

8.1 Introduction

Endogenic processes in geology are a function of a body's internal geodynamic activity. They comprise volcanic, tectonic, and isostatic processes, which shaped the surfaces of all terrestrial planets, the Moon, and basically all other Solar System bodies with solid surfaces that have been observed in some detail. The most recent spacecraft observations have confirmed this notion, and revealed past or present endogenic activity even on bodies where this was not previously expected (Fig. 8.1). The study of endogenic processes and their resulting landforms and landscapes puts important constraints on the internal evolution (Chap. 10) and the surface history of a geologic body (Chap. 11).

This chapter focuses on rather large-scale surface features (10^0–10^6 m). Although rovers and landers can study much smaller features, they are few in numbers and are typically very restricted in their operational range (Chap. 5), although they did provide invaluable information on the magmatic evolution of their target bodies. Another important source of information, specifically on the petrology of magmatic

E. Hauber (✉)
German Aerospace Center (DLR), Berlin, Germany
e-mail: ernst.hauber@dlr.de

D. Mège
Space Research Centre, Polish Academy of Sciences, Warsaw, Poland
e-mail: dmege@cbk.waw.pl

T. Platz
Max Planck Institute for Solar System Research, Göttingen, Germany
e-mail: platz@mps.mpg.de

P. Brož
Institute of Geophysics, Czech Academy of Science, Prague, Czech Republic
e-mail: petr.broz@ig.cas.cz

A.P. Rossi, S. van Gasselt (eds.), *Planetary Geology*, Springer Praxis Books,
DOI 10.1007/978-3-319-65179-8_8

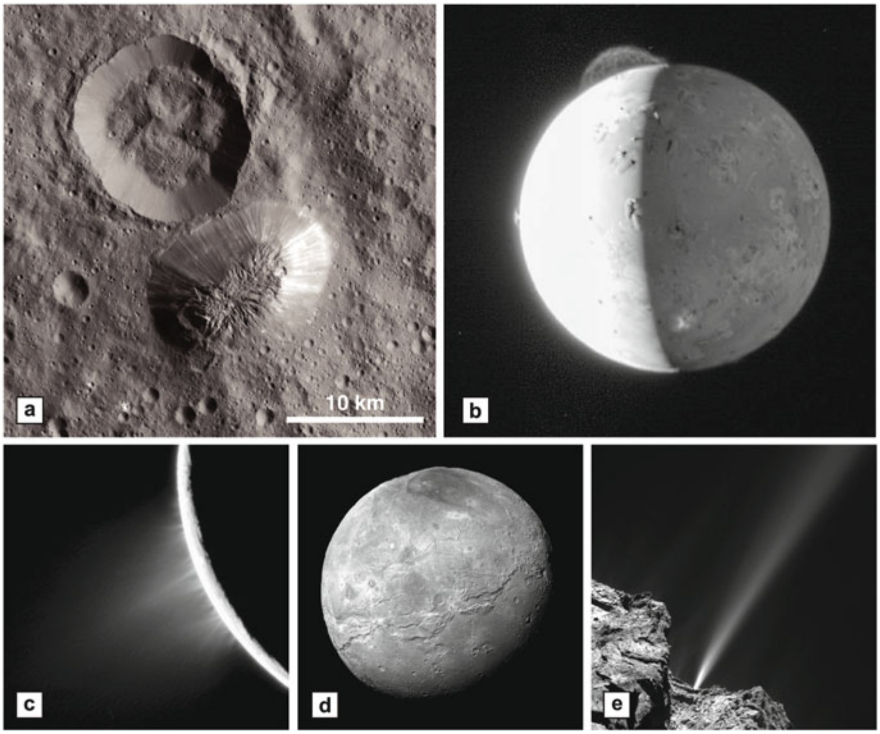

Fig. 8.1 Evidence for endogenic activity on small bodies beyond the terrestrial planets. While some of these processes were predicted on theoretical grounds (e.g., on Io) or expected from Earth-based observations (e.g., on comets), other endogenic activities came as a surprise (e.g., on Enceladus or Pluto and Charon). (**a**) A potential cryovolcanic dome on the dwarf planet Ceres. The conical edifice has a basal outline of about $\sim 10 \times 20$ km and stands about 5 km tall above its surroundings. (**b**) A plume over the active volcano, Tvashtar, on Io. The plume reaches a height of 290 km above the surface. It was imaged on 28 February, 2007 by the New Horizons spacecraft on its way to Pluto. (**c**) Jets of ice particles, water vapor and trace organic compounds emanating from the surface of Enceladus. These *ice geysers* on the ∼475 km-diameter satellite of Saturn were detected by the Cassini spacecraft. (**d**) Pluto's satellite, Charon, displays a surprisingly varied and partly young surface. The prominent tectonic fractures are evidence for stresses acting on its brittle outer shell. (**e**) A short-lived outburst from comet 67P/Churyumov-Gerasimenko. The jet is thought to have a speed of at least 10 m/s (see also Chap. 13). Source: (**a**) NASA/JPL-Caltech/-UCLA/MPS/DLR/IDA. (**b**, **d**) NASA/Johns Hopkins University APL/SRI. (**c**) NASA/JPL/SSI. (**e**) ESA/Rosetta/MPS for OSIRIS Team MPS/UPD/LAM/IAA/SSO/INTA/UPM/DASP/IDA

rocks, comes from meteorites (Chap. 6), e.g., from Mars or asteroids: All such studies need to be complemented by modeling to arrive at a physical understanding of the driving forces and mechanisms.

The study of endogenic processes addresses some of the most fundamental questions in planetary geology, including the bulk composition, the history of accretion and differentiation, the heat generation and transport, and the evolution of planetary atmospheres (via outgassing) and climates (Chap. 10). Igneous intrusive and extrusive magmatic processes operated or still operate on all terrestrial planets, possibly on larger asteroids or their parent bodies, and on some satellites in the outer Solar System, and tectonic processes left their traces on all observed objects. We first introduce the inventory of tectonic and volcanic landforms on the various planetary bodies, and present techniques how their morphology can be analyzed. We then discuss how tectonic regimes differ between the individual bodies, and how magmatism evolved on them. Given the enormous range in number and variety of endogenic processes in the Solar System, this chapter necessarily needs to limit its scope to a few major aspects. More detailed treatments of the geology of the icy satellites and minor Solar System bodies are given in Chaps. 12 and 13, respectively. Impacts can trigger significant tectonic movements and the generation of impact melt (Chap. 7).

8.2 Landforms of Endogenic Processes

8.2.1 Tectonic Landforms

The mechanical deformation of planetary lithospheres produces tectonic structures such as joints, deformation bands, faults and folds. Therefore, the study of tectonic features (*structural geology*) can help to reconstruct the history of crustal and lithospheric stresses and their causative factors, constraining models of geologic evolution and internal processes (see Chap. 10). Tectonic structures are observed on all larger Solar System bodies with a solid surface. There is no fundamental difference in the main structural elements between the Earth and most other Solar System bodies. This is not very surprising, as both the basic constituents of the deformed solids (i.e., rocks and ice) and the mechanical stresses responsible for deformation are similar throughout the Solar System. Nevertheless, the relative abundance of tectonic surface features varies widely among the planets, their satellites and asteroids. This can mostly be explained by the nature of the stresses and the tectonic style operating on different bodies. For example, strike-slip faulting is characteristic for the horizontal movement of lithospheric plates against each other and is a hallmark of plate tectonics. As Earth is the only planet where plate tectonics operate at present, it is not surprising, then, that large-scale strike-slip faulting is rare on one-plate planets such as Mars, Mercury, or the Moon.

Fractures are arguably the most common structural elements on planets and small bodies. They are typically classified into mode I fractures (opening mode: relative motion is normal to the fracture plane), mode II fractures (shear mode: motion is parallel to the fracture plane), and mode III fractures (tear mode: motion is parallel to the tip line of the fracture). Typical examples of mode I fractures are joints or tension cracks (Fig. 8.2). Although they are typically too small to be identified in remote sensing data, joints have been observed on Mars in very high-resolution images. Joints can control the preferred orientation of large-scale

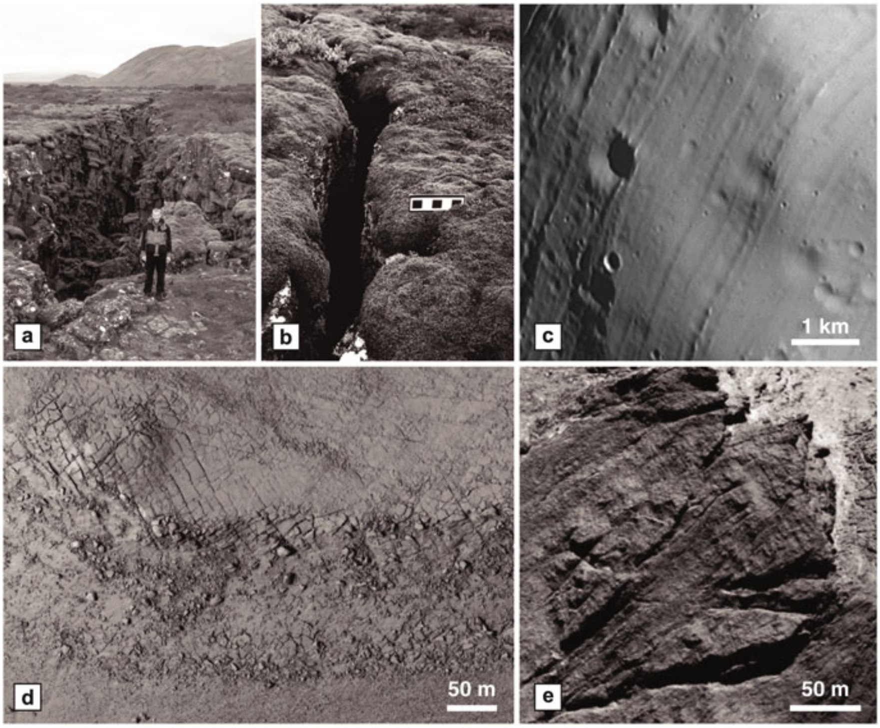

Fig. 8.2 Joints and tension cracks on Solar System bodies. (**a**) Tension fractures in the rift zone at Thingvellir, Iceland. (**b**) Close-up of tension fractures at Thingvellir: *Squares* on scale bar in are 5 × 5 cm. (**c**) The origin of grooves on the Martian satellite, Phobos, is unknown, but some studies favour a formation as tension cracks. (**d**) Joints in rocks near the Nilosyrtis Mensae region, Mars. (**e**) Fractures on comet 67P/Churyumov-Gerasimenko. Source: (**a**, **b**) E. Hauber. (**c**) ESA/-MEX/DLR/FU Berlin. (**d**) NASA/MRO/HiRISE/University of Arizona. (**e**) ESA/Rosetta/MPS for OSIRIS Team MPS/UPD/LAM/IAA/SSO/INTA/UPM/DASP/IDA

erosional structures such as yardangs, hence the orientation of joint patterns can potentially be inferred from the analysis of such erosional landforms. On a much larger scale, it has also been hypothesized that some obviously tectonic fractures at the giant Valles Marineris system of linear troughs on Mars are tension fractures. Sets of linear grooves that have been detected on the Martian moon, Phobos, and on other irregular small bodies (e.g., the asteroid 21 Lutetia) may also be opening fractures, although a variety of models has been proposed to explain their origin. Most recently, new images of comet 67P/Churyumov-Gerasimenko revealed ubiquitous cracks that may have been formed as opening fractures by internal (or thermal) stresses (Chap. 13).

Mode II fractures are faults, which can further be subdivided into normal faults, reverse or thrust faults, and strike-slip faults. Normal faults are records of tensional stresses and extensional deformation, whereas reverse and thrust faults indicate compressional stresses and contractional deformation.

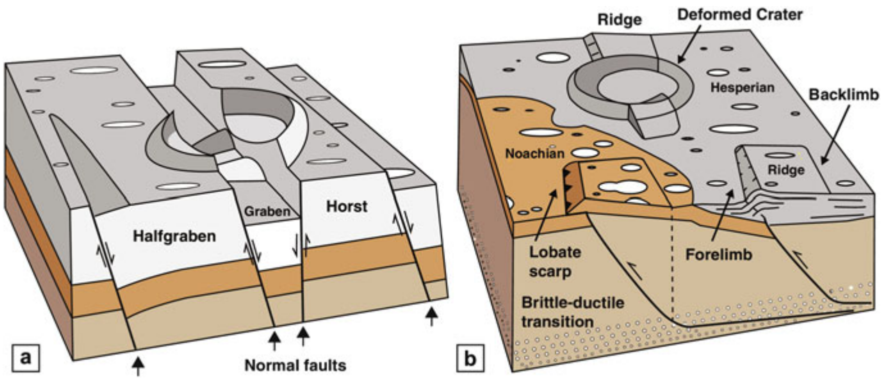

Fig. 8.3 Schematic views of major fault types on the terrestrial planets. (**a**) Extensional features. (**b**) Contractional features. Source: redrawn after Mueller and Golombek (2004)

Normal faults cut the surfaces of most, if not all planetary bodies. They are typically visible as rectilinear scarps (Fig. 8.3). Due to the lack of vegetation and the low erosion rates, fault scarps on, e.g., Mars or the Moon appear typically very pristine and can often be better studied in remote sensing data than their terrestrial counterparts. The maximum lengths of normal faults on Mars and Venus can be tens to hundreds of kilometres, although the longest faults are often composed of several linked segments. As on Earth, planetary normal faults seem to grow by segment linkage, and *en echelon* patterns and relay ramps are common. On the short end of fault lengths, they can be observed down to the resolution limit of images. Similarly, fault throw, though difficult to determine in remote sensing data, ranges from meters to kilometres. The throw is typically estimated from the vertical scarp height and an assumed fault dip. Based on the available topographic information, it seems that the displacement-length relationship of planetary normal faults is similar to that on Earth. The fault throw derived with this method can be used to constrain the extension and strain across the fault scarp. As mentioned above, these estimates all require the use of an assumed fault dip, which is typically chosen to be 60°. Based on analogy to the Earth, this assumption relies on the classical dip value for normal fault planes. It is known, however, that this value, which is based on physical principles, applies to faults at a certain depth, whereas normal faults commonly start as subvertical tension fractures and are much steeper than 60° near the surface. Unfortunately, the true dips of normal fault planes are difficult to determine from remote sensing data. Visible scarps are typically not the actual fault planes, but represent talus deposits generated by erosion and fault scarp retreat. Therefore, the measurable slope often corresponds to the angle of repose and cannot be used to constrain the true dip.

Normal faults can occur as isolated features, but more commonly they are part of a more complex structural architecture. Half-grabens (asymmetrical) and grabens (symmetrical) have been observed on the Moon, Mars, Venus, and recently also on Mercury and asteroids. They typically form sets or families, which bear records

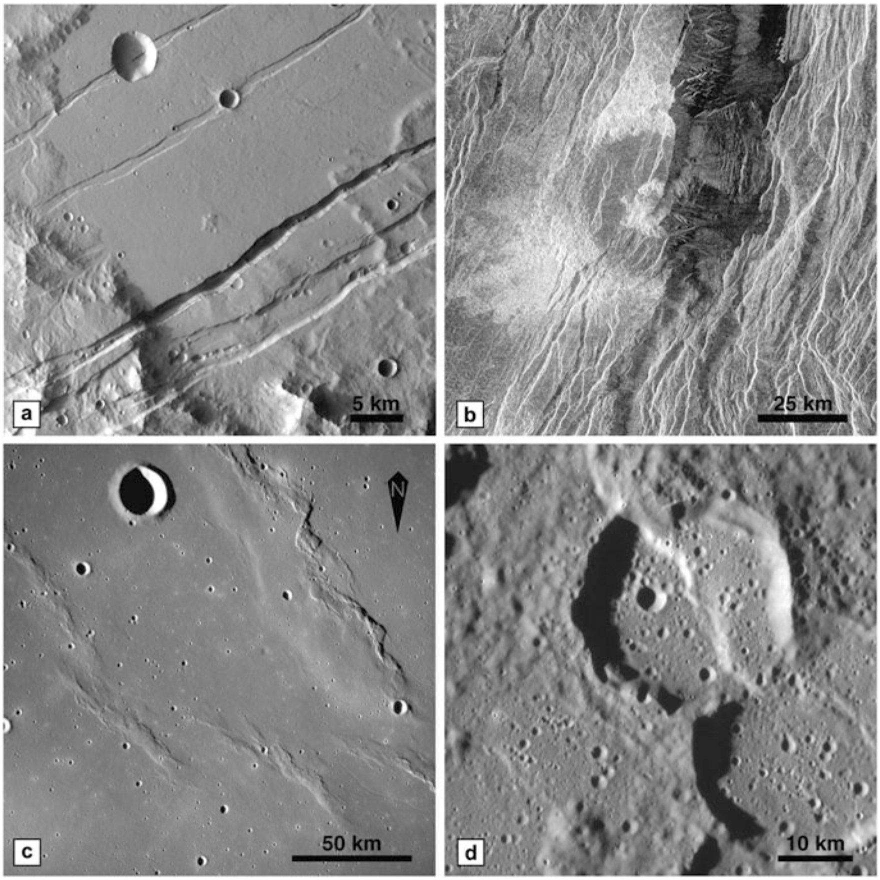

Fig. 8.4 Examples of main types of planetary faults. (**a**) Detail of a set of long and narrow grabens in the Memnonia region of Mars. North is up, illumination from left (west). (**b**) Large-scale extensional rift between Rhea and Theia Montes in Beta Regio on Venus as seen in Magellan radar data. Note rifted crater with a diameter of ∼37 km. (**c**) Lunar wrinkle ridges north of Flamsteed crater, Oceanus Procellarum. (**d**) Lobate scarp in the Rembrandt basin on Mercury. Source: (**a**) ESA/MEX/DLR/FU Berlin, HRSC orbit 4073. (**b**) NASA/JPL. (**c**) Apollo 12 image, NASA. (**d**) NASA/Johns Hopkins University APL/Carnegie Institution of Washington

of local- or regional-scale processes (Figs. 8.4 and 8.5). For example, concentric grabens along the margins of lunar maria were formed by the relatively larger subsidence in the mare interiors caused by volcanic loading (Chap. 11).

Another prominent example of planetary graben sets are the long and narrow grabens that radiate outwards from several magmatic centers in the Tharsis region on Mars, the largest known volcano-tectonic province in the Solar System.

Even more complex extensional systems resemble terrestrial continental rifts. They are only observed on larger terrestrial planets, i.e., on Mars, Venus and Earth. Venusian rifts in particular are analogous to rifts in, e.g., East Africa, and

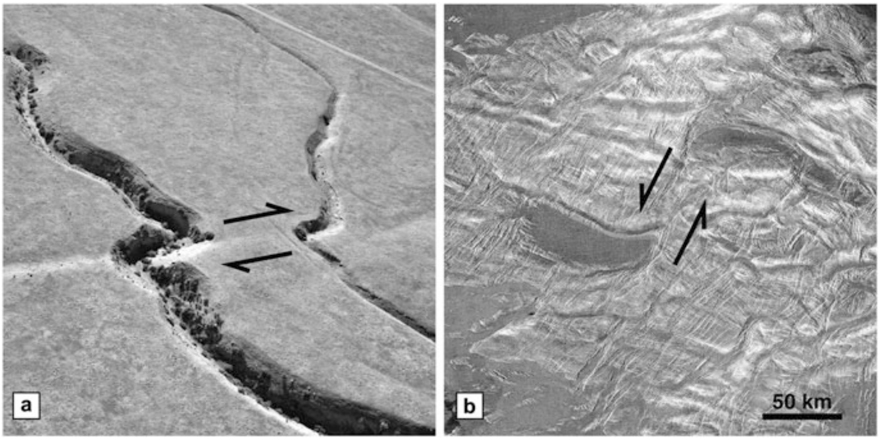

Fig. 8.5 Examples of main types of planetary faults. (**a**) Strike-slip fault as part of the San Andreas graben in California. Fault section length is ~150 m. (**b**) Strike-slip fault in Ovda Regio on Venus. Source: (**a**) USGS photograph by David K. Lynch, Kenneth W. Hudnut and David S.P. Dearborn (2009). (**b**) NASA/JPL/Magellan

are obviously associated with major volcanic centres. Rifts on Mars and Venus display a number of properties that are characteristic for terrestrial rifts, such as rift segmentation, segment linkage at accommodation zones, and alternating rift polarity, i.e., an asymmetric cross section with a dominating master fault system at one rift margin, which changes position at accommodation zones. Extension across planetary rifts appears to be limited to a few kilometres, and strain is accordingly low, too. Terrestrial continental rifts are believed to penetrate the entire lithosphere, and indicate the beginning of continental break-up. The rifts on Mars and Venus, though similar in dimension and style, are not thought to be evidence for plate tectonics, but they are obviously major structural elements and could indicate an early stage of large-scale tensional strain localization which never developed towards a more advanced stage.

Contractional structures are widespread on all terrestrial planets. The most commonly observed class of contractional landforms are wrinkle ridges (Fig. 8.4). This descriptive term was chosen because they are characterized, in cross section, by a broad topographic rise or arch on which is superposed a smaller and steeper ridge. On the top of this ridge, an even smaller crenulation has an irregular (*wrinkled*) trace in plan-view. The ridge is typically offset to one side of the broad arch, creating an asymmetric profile with a steeper forelimb and a shallower backlimb (Fig. 8.3). The width of wrinkle ridges is variable, ranging from a few kilometres to few tens of kilometres, and the maximum height is typically a few hundred meters. Although several models have been invoked to explain wrinkle ridges, the most accepted view holds that they are thrust-propagation folds, where near-surface layered strata are folded by the motion along a blind thrust at depth (a blind thrust does not propagate to the surface and thus does not break it). Since the first wrinkle ridges studied in detail were those on the volcanic lunar mare plains, it was once thought that they are indicative of a substrate consisting of layered basaltic lava flows. However, at least in the case of Mars, it cannot be excluded that sedimentary layers may also be folded above blind thrusts. Sometimes the plains to one side of the ridge are systematically elevated with respect to those on the other side, thus indicating the vergence of the thrust. The depth to which wrinkle ridges extend into the crust is a matter of debate. The *thin-skinned* models favour a limited depth of faulting, perhaps originating at a décollement that may correspond to a rheologic boundary layer (e.g., the brittle-ductile transition). On the other hand, a *thick-skinned* origin would involve faulting that penetrates perhaps most of the crust. Wrinkle ridges frequently display an evenly spaced pattern in plan-view, with distances of typically tens of kilometres between individual ridges. Good examples of such regular patterns can be found in the plains north and south of Valles Marineris on Mars. This spacing has been used to model the depth of the deformed layer, with shorter distances between wrinkle ridges corresponding to a thinner volume. The origin of wrinkle ridges may often be related to loading and resultant compressional stresses. For example, wrinkle ridges on the lunar mare plains are thought to result from subsidence in the mare centres due to the load of the basaltic infilling, and corresponding compressional stresses acting on the near-surface crust.

The second main class of planetary contractional features are lobate scarps. As typical for the nomenclature of planetary landforms, this term is descriptive and avoids interpretation. Lobate scarps are mostly observed on the Moon and Mercury. The characteristic landform is a topographic scarp that is lobate in plan-view (Fig. 8.3). The commonly accepted explanation is that lobate scarps are surface-breaking thrust faults. Typically, it is assumed from rock mechanics that the fault planes of lobate scarps have dips of $\sim 30°$. The topography and geometry of lobate scarps can yield clues to fault motion or lithospheric properties. The cross-sectional geometry of a lobate scarp, especially the height and the location of a trailing syncline, can be used to model the thickness of the layer that has been bent upwards by the motion along the fault plane. As this thickness is assumed to be equal to the thickness of the elastic lithosphere, this method enables estimating the heat flow at the time of faulting. The best-known example how lobate faults can be used to constrain planetary evolution is illustrated by studies of Mercury's cooling history. Mapping of the global lobate scarp population reveals their number and their individual lengths. Together with a given ratio of displacement to length, and an assumed fault dip, it is possible to calculate the total amount of cumulative crustal shortening (i.e. loss of area) that was accommodated by the lobate scarps. This surface shrinkage can then be used to estimate an associated decrease of the planet's radius.

Except on Earth (Fig. 8.5), strike-slip faulting is much less common than normal or thrust faulting on bodies with rocky surfaces: Only Venus displays some large-scale strike-slip faults (Fig. 8.5), whereas their existence on Mars is contentious. This positive correlation of increasing planetary mass with increasing importance of strike-slip faulting is not surprising, as lithospheric mobility and relative lateral motions on planets without plate tectonics (*one plate-planets*) are very limited and basically restricted to local-scale processes. However, some small-scale strike-slip movement may have occurred in wrinkle ridged-plains on Mars and Mercury.

As compared to the ubiquitous brittle deformation, the observational evidence for ductile deformation is much sparser in the Solar System. Folding of rocks appears to be more common on Earth and Venus as compared to Mars, Mercury, and the Moon. As ductile deformation and folding occurs in deeper levels of the crust and lithosphere and is only exposed at the surface after uplift and erosion, it is typical for mountain belts, e.g., in the orogens created by plate collisions on Earth. As erosion rates are much lower on the smaller planets due to the lack of dense atmospheres, the exposure of deeper crustal levels is rare, and except on Earth, large folds have only been observed on Venus. On a smaller scale, folding due to sedimentary processes (e.g., soft sediment deformation) is apparent in high-resolution images of sedimentary rocks on Mars.

Apart from these landforms, which are all not restricted to any given planet, there are more special features that are unique to some bodies. Among the terrestrial planets, this applies especially to Venus, where a number of large-scale volcano-tectonic landforms may be related to diapirism (both upwelling and downwelling processes have been proposed). The best-known examples are the coronae on Venus, which are circular or slightly elliptical in plan-view, have diameters of up to a few hundred kilometres, and are commonly associated with concentric fracture zones and one to several volcanic centres. Other unique features on Venus are novae and arachnoids, which also reach significant dimensions, but are characterized by radial tectonic fractures. Io, Jupiter's volcanically active satellite, shows some very large mountains that seem to be uplifted along deep-seated thrust faults. The compressive forces at depth are thought to be a result of volcanic loading at the surface, which causes downward movement of the deeper crustal levels but no contraction at the surface.

8.2.2 Volcanic Landforms

Terrestrial magmatic processes produce a variety of landforms over a broad range of scales. Most of these landforms have morphological analogues on other terrestrial planets and the Moon (Fig. 8.6). The study of volcanic landforms can inform the analysis of eruption processes and, therefore, enables a better understanding of the chemical and physical characteristics of magmatic systems. Typically, volcanic landforms are classified on the basis of magma composition and the types of eruptions (e.g., effusive vs. explosive) and their products, but such classifications should also consider additional parameters such as geomorphic scale, constructional vs. erosional origin, and mono- vs. polygenesis. It is important to note that volcanic landforms are not restricted to volcanoes, but also to eruption products that may have been emplaced far from any vent (e.g., lava flow fields, ash fall deposits).

Magmatism is strongly coupled to tectonism; hence volcanic landforms are typically not randomly distributed over a planet's surface and rarely occur isolated.

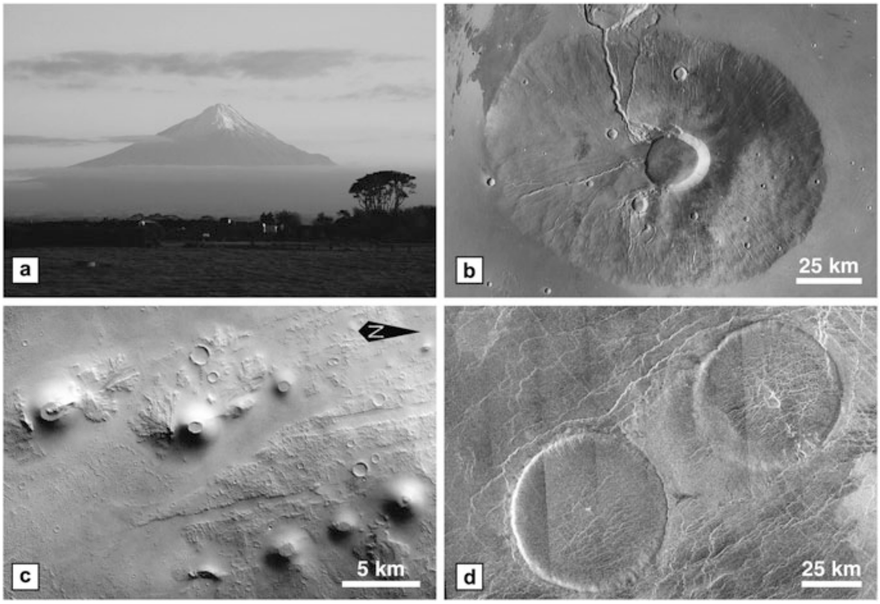

Fig. 8.6 Examples of volcanic edifices. (**a**) Mt. Taranaki (New Zealand) is an andesitic composite volcano rising from sea level to 2518 m. The summit crater hosts remnants of a lava dome. (**b**) Ceraunius Tholus located in Mars' Tharsis volcanic province is a large shield volcano. It is partially buried by the surrounding lava plains. The summit is marked by a near circular caldera. (**c**) A complex of pyroclastic cones is located in the Ulysses Fossae area north of Biblis Tholus, Mars. (**d**) Two steep-sided, flat-topped volcanic domes located in Tinatin Planitia, Venus, are shown on this Magellan radar image. They formed by extrusion of highly viscous lava. The largest dome is 62 km across. Source: (**a**) T. Platz. (**b**) ESA/MEX/DLR/FU Berlin. (**c**) NASA/MRO/CTX. (**d**) NASA/JPL/Magellan

Instead, their spatial distribution reflects the nature of endogenic processes that produce magmatism. On Earth, most magmatic activity is associated with plate tectonics, and the majority of volcanic landforms are concentrated at divergent and convergent plate boundaries. However, there are also huge volcanic and plutonic deposits and provinces that lie outside tectonic plate boundaries. They often form extended continental flood basalts and oceanic plateaus, and are collectively termed Large Igneous Provinces (LIPs). LIPs are typically characterized by effusion rates that are much higher than those observed at divergent plate margins, and are important because they enable studying lithospheric and mantle processes that are often obscured by plate tectonics. It is commonly assumed that LIPs form when the head of a mantle plume impinges on the crust and large amounts of magma are emplaced as intrusions and basaltic lava flows (but not all researchers agree on the link of intraplate volcanism to plumes). On other planetary bodies, the distribution of volcanic surface features seems to be related to narrow or broad sites of mantle upwelling. For instance, the vast volcanic plains on the Moon and Mercury are concentrated at the nearside and at a few basins and the North Polar Region, respectively. On Mars, volcanism has been active over most, if not all of its history. Whereas older (∼3.9 Ga) volcanic features are globally widespread, Amazonian volcanism (<3.4 Ga) has basically focused on two distinct large provinces, Tharsis and Elysium.

At smaller scales, volcanoes commonly tend to form clustered volcanic fields. Hundreds of individual edifices can be clustered in a single volcanic field, and the analysis of their spatial distribution can reveal structural trends that controlled their emplacement. If an individual edifice was formed during one single eruptive cycle, it is called a monogenetic volcano, and accordingly a monogenetic volcanic field consists of many monogenetic volcanoes. A typical example for a monogenetic volcano is a scoria cone, and actually scoria cones are the most common volcano type on Earth. In contrast, polygenetic volcanoes (e.g., composite volcanoes) form over several eruptive cycles.

Calderas are volcano-tectonic structures common on both Earth and Mars: they represent collapsed roofs above shallow magma chambers that were partly emptied: such landforms are rather common on Mars on various volcanic edifices, including very large ones, e.g., at the top of Olympus Mons (Fig. 8.7).

As for the Earth, planetary volcanic landforms are frequently classified according to the associated eruption type (Table 8.1). Most authors distinguish only a handful of eruption types, typically as a function of the way how the gas exits the magma. If magma ascends quietly to the surface and is not significantly disrupted by gas release, an *effusive eruption* produces lava flows. They can pile up to form mostly basaltic shield volcanoes such as on Hawaii, but more importantly in the geological past, such eruptions also contributed to the formation of huge flood basalt provinces,

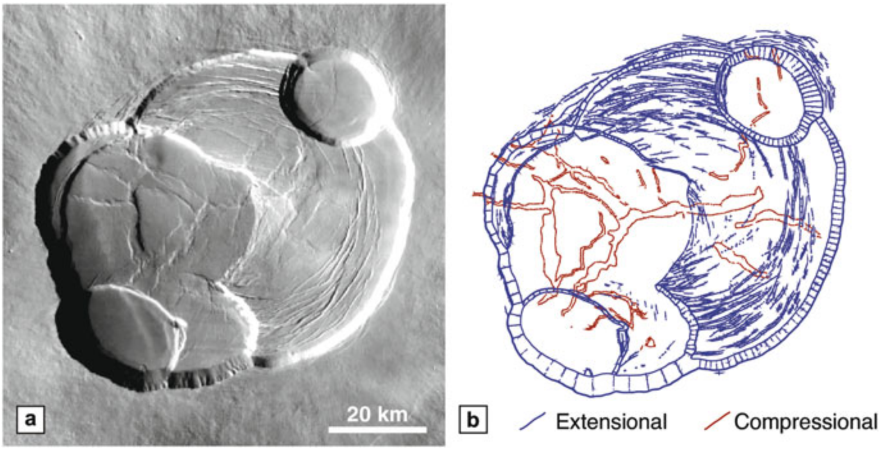

Fig. 8.7 Volcano-tectonic features at Martian volcanoes. (**a**) The summit of Olympus Mons (height: 21.2 km) is characterised by a nested caldera complex. (**b**) Structural map of extensional landforms (normal faults, grabens) on the caldera floor, including both extensional and contractional features. Source: (**a**) ESA/MEX/DLR/FU Berlin. (**b**) P. Kronberg and E. Hauber

Table 8.1 Volcanic eruption types (after Green and Short (1971), Wilson and Head (1994), Sigurdsson (2000))

Eruption name	Type	Description
Icelandic	Effusive	Fissure eruptions (gas-poor) producing great volumes of basaltic lava flows. Forms plateau basalts.
Hawaiian	Effusive	Eruptions from fissures, calderas, and pits; low-viscosity lavas that are erupted quietly, but can display some moderately explosive behaviour (lava or fire fountaining). Fire fountains erupt clods of lava that fall back still in molten state, form a lava pond, and continue as lava flow.
Strombolian	Explosive	Moderately explosive, rhythmic or continuous explosions; gas rises faster than magma in the conduit, bubbles form and coalesce until they disrupt the lava; ejection of lava clods that travel in ballistic trajectories and cool rapidly, falling down as bombs and unconsolidated scoria. Typical landforms are scoria cones.
Vulcanian	Explosive	Small to moderate-sized, short-lived (seconds to minutes) volcanic outbursts that eject material to heights <20 km. Discrete, violent explosions and ballistic ejection of blocks and bombs. More viscous lavas than in Strombolian eruptions can form a crust, under which gas pressure can build up until it is released violently.
Peléan	Explosive	Slow rise of high viscosity lava forms viscous dome or plug with solid carapace, catastrophic disruption, gas (and some lava) escapes from lateral flank openings, or destruction of plug; origin of a small convecting plume and pyroclastic flow (nuées ardentes) that can create block-and-ash flow deposits, made up from lava blocks with fine vesicles.
Plinian	Explosive	Magma disrupts into small fragments, efficient heat transfer to gas, entrainment of fine particles into rising gas stream, entrainment of ambient atmosphere; formation of very high convective ash plumes; creates (wind-transported) ash fall deposits, or, if the rising gas column becomes too dense and collapses, pyroclastic flow deposits (ignimbrites or ash-flow tuffs).

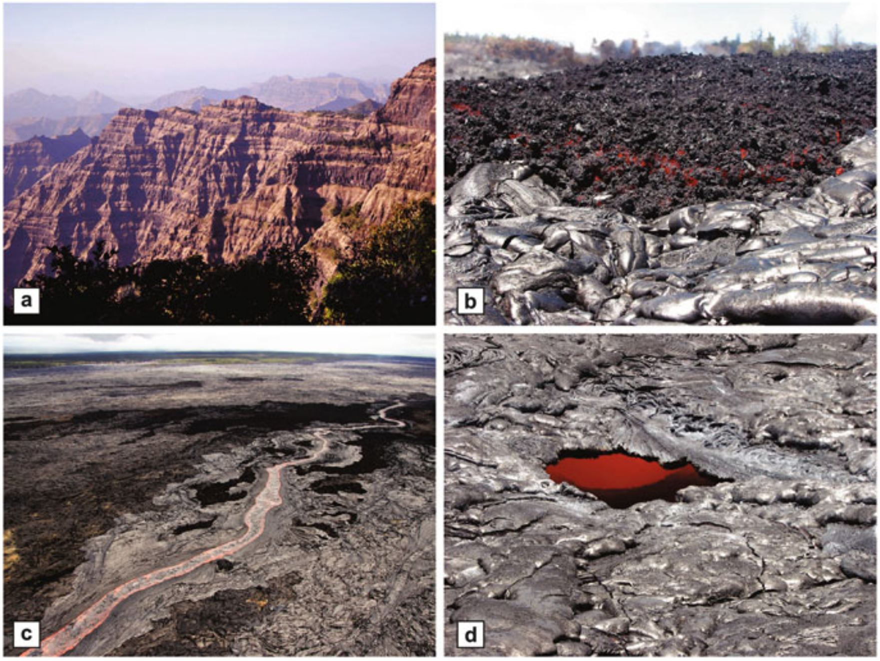

Fig. 8.8 Lava flow morphologies. (**a**) Trapp basalts in the Deccan province, India. Multiple flat-lying lava flows were eroded into mountains with a characteristic step-like topography. (**b**) An active *'a'ā* flow (background) advances on top of a recent *pāhoehoe* flow (foreground) in the March 2008 eruption of Kilauea, Hawaii. (**c**) Channelised lava flow on the northeastern flank of Pu'u 'Ō ō, Hawaii. (**d**) Skylight on a tube-fed lava flow on Pu'u 'Ō ō, Hawaii. Source: (**a**) Gerta Keller. (**b**) USGS. (**c**) Hawaiian Volcano Observatory, USGS. (**d**) USGS

e.g., the Deccan Traps in India or the Siberian Traps (the word trap comes from the Swedish term *trappa* for stairs and refers to the characteristic step-like topography of sub-horizontally layered flood basalts stacks (Fig. 8.8)). Several properties of lava flows have been used for their subdivision into separate types. The perhaps best known is the distinction between *pāhoehoe* flows and *'a'ā* flows, which may have identical chemical compositions yet display smooth and ragged surface textures, respectively (Fig. 8.8). It is commonly assumed that lava rheology, strain rate, and effusion rate govern whether a *pāhoehoe* or *'a'ā* flow is being emplaced.

Importantly, a single flow can transition from *pāhoehoe* type to *'a'ā* type, but not vice versa. Another morphologically relevant distinction is that between *channelised flows* and *tube-fed flows*. Open-channel flows typically display a central channel and lateral levées (Fig. 8.9), whereas tube-fed flows form an upper cooling crust, while the molten lava continues to flow underneath the crust. When this crust collapses, a so-called *skylight* (Figs. 8.8 and 8.9) is formed, and when the tube drains due to a ceasing lava supply, it turns into an empty lava conduit (often called lava tunnel). Tube-fed flows cool less efficiently than channelized flows due to the

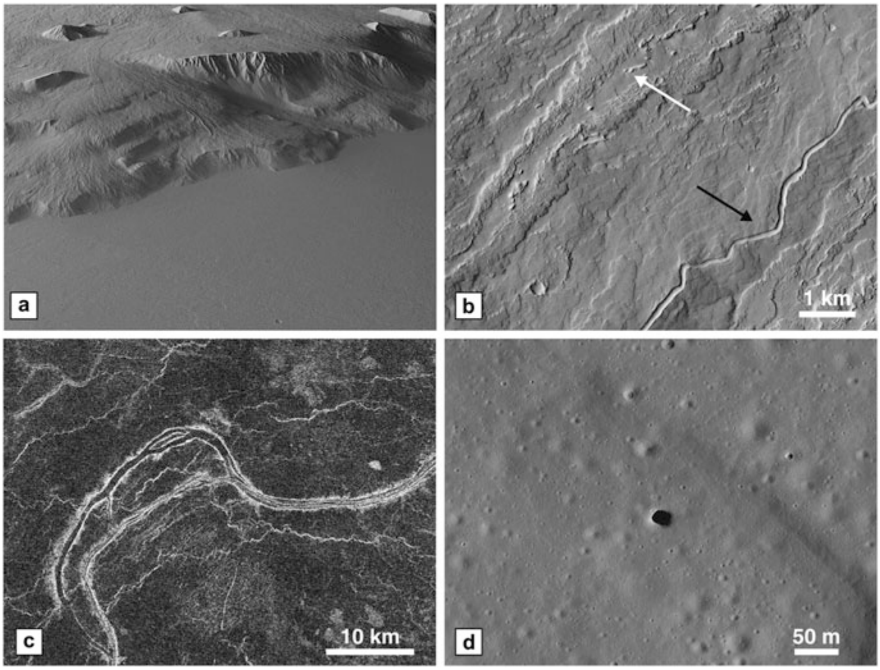

Fig. 8.9 Lava flow fields on the terrestrial planets. (**a**) Perspective view of the southeast flank of Olympus Mons. Successions of lava flows cascade down a 5-km high scarp. The lava fan at the base of the scarp is truncated by an extended smooth lava plain (foreground). Scene is about 70 km across. (**b**) The lava plain approx. 30 km NNE of Olympus Mons shows different lava flow types. *Lower right*: sinuous channel atop a lava tube, indicated by the *black arrow*; *upper left*: up to 2 km wide lava flow with characteristic levée and channel facies (*white arrow*). (**c**) Lava channel in Sedna Planitia, Venus. (**d**) Orbital view of a skylight on a lava-covered plain the Moon, see Fig. 8.8. Source: (**a**) ESA/MEX/DLR/FU Berlin, HRSC orbit 11524. (**b**) NASA/MRO/CTX, modified after Platz et al. (2015). (**c**) NASA/NSSDC/Magellan. (**d**) NASA/LRO/LROC/GSFC/ASU

protective effect of the solid crust, which minimizes radiative and convective heat loss. The cooling rate has an effect on lava flow length, and a lava flow stops when the heat loss coupled with increases in the crystal content force the viscosity and yield strength to increase to a point where motion is no longer possible (*cooling-limited flows*). Alternatively, lava flows can come to a halt before the maximum (cooling-limited) extent is attained when the lava supply stops (*volume-limited flows*). Composition, insulation and morphology being constant, flow length will positively correlate with effusion rate.

If the pressure release associated with the magma ascent leads to volatile exsolution and the formation of gas bubbles, these bubbles can ultimately disrupt the magma, leading to an *explosive eruption*. Depending on the nucleation depth of gas bubbles in the conduit, the viscosity, the ascent rate, and other factors, explosive eruptions can be more or less violent. Accordingly, several types of explosive eruptions are distinguished, ranging from moderately explosive Hawaiian eruptions to highly explosive and potentially hazardous Plinian eruptions. Each eruption type produces certain landforms, although the distinction in remote sensing data is not always straightforward.

The interaction of magma with water and/or ice leads to rapid heating and water vapour formation. If this happens in a contained environment (e.g., groundwater heating in the subsurface), the vapour pressure can increase sufficiently to disrupt the containing material, leading to extremely violent *phreatomagmatic eruptions*.

Not all eruptions in aqueous environments are violent, however. Basaltic eruptions at mid-ocean ridges are typically quiet, because the high ambient hydrostatic pressure prevents violent magma degassing. A typical product of such non-violent subaqueous eruptions are pillow lavas, which may represent the most abundant volcanic rock type on Earth and are characterized by rounded (pillow-like) shapes and a rapidly quenched glassy rind. Other typical volcanic landforms on the ocean floors are individual volcanoes called seamounts. Subglacial eruptions represent a special type of magma-ice/water interaction. When the volcanic heat melts the overlying ice, a subglacial lake is formed and the eruption becomes subaqueous and may form pillow lavas first (if the lake is deep enough) and more explosive eruption products in shallow water later.

It is important to note that the style of an eruption can (and typically does) change during a single eruptive cycle. For example, an eruption can evolve from explosive to effusive or vice versa, and eruptions can change from an initial *wet* phreatomagmatic stage to a subsequent *dry* (e.g., Strombolian) explosive style. Accordingly, landforms created during an eruption may reflect these varying eruptive conditions, and any interpretation based on remote sensing data needs to acknowledge that the resulting landform assemblage may have a polyphase eruption history, and needs to consider the possibility of mixed deposits.

8.3 Tectonism: Driving Forces

Forces generate stress that generates deformation. In planetary crusts or lithospheres, contact forces, which act on rock surfaces, and body forces, which affect the entire volume of rock, coexist. Contact forces can be generated by a number of ways; for instance, on Earth, the weight of the cooling oceanic crust makes it enter into subduction zones, a process which generates traction forces (slab pull). Those forces generate the global system of stresses that drive the lithospheric plates and produce orogenic belts, rifts, and large strike-slip faults. On the other hand, body forces that can result in rock deformation are generated by gravity. Landsliding is an example of gravity tectonics.

Typically, the tectonic style is thought to be identical to the crustal deformation style. However, this style may not depend on the crust only. What is instead relevant is the portion of the planetary radius located above the thermal boundary layer above the convective mantle, if any. On Earth, this zone includes the uppermost part of the mantle and the crust, of distinct composition, which define the mechanical lithosphere. Alternatively, and more simply, the lithosphere may also be defined as the region in which heat is transported to the surface mainly by conduction. Both lithosphere definitions give similar thickness estimates, though calculated differently. Another useful definition of lithosphere is based on the speed of seismic wave propagation. Whatever the definition on other planets, the absence of deep geophysical data is a major problem.

The tectonic style in the lithosphere varies with time, following the evolution of the heat flow, as high heat flow promotes ductility. It is, for example, not clear when plate tectonics started on the early Earth. On the other hand, Venus may have experienced plate tectonics in the past for which we do not see evidence anymore. In addition, a lithosphere that does not deform instantaneously may deform on geologic timescales by viscous relaxation. The style of deformation will then depend as well on the lithospheric composition, the water content, and the loading conditions. On Earth, where the lithosphere is made of two compositionally different components, crust and mantle, the question of how continental lithosphere behaves on the long term in response to these factors is not yet clear, in spite of the wealth of data available. Understanding the style and origin of tectonic activity of other planets is therefore a formidable challenge, in the absence of data regarding the composition and mechanical properties of the lithosphere, and taking into account the huge uncertainties on how the heat flow and water content of the crust have evolved through time.

8.3.1 The Tectonic Style of the Earth

Plate tectonics is the best documented example of global tectonics (Fig. 8.10). It currently operates on Earth only and requires several conditions. One is the presence of volatiles in the crust, which facilitates plate overthrusting in subduction zones, and for which evidence is testified since the Archean. Another one is that the plates are rigid and can deform at their margins only. The driving force is slab pull at subduction zones, but at smaller scale, body forces provided by density contrasts, for instance due to magmatic underplating, can influence intraplate stress conditions and hence the tectonic style. Although the view that plates are generally decoupled from the underlying mantle is frequently shared, it is increasingly considered that mantle flow contributes significantly to plate motion beneath oceans and kinematics at convergent plate margins.

The demonstration that plate tectonics operates on Earth was enabled combining information from two critical types of measurements. The symmetric distribution of magnetic anomalies in rocks of increasing age away from the mid-ocean ridges requires seafloor spreading on geologic timescales. GPS (Global Positioning System) and interferometric methods show that instantaneous plate displacement rates are equal to the displacement rates measured on geologic timescale, and demonstrate that internal plate deformation is negligible and that strain concentrates at sharp to diffuse boundaries. It is important to note that both methods require

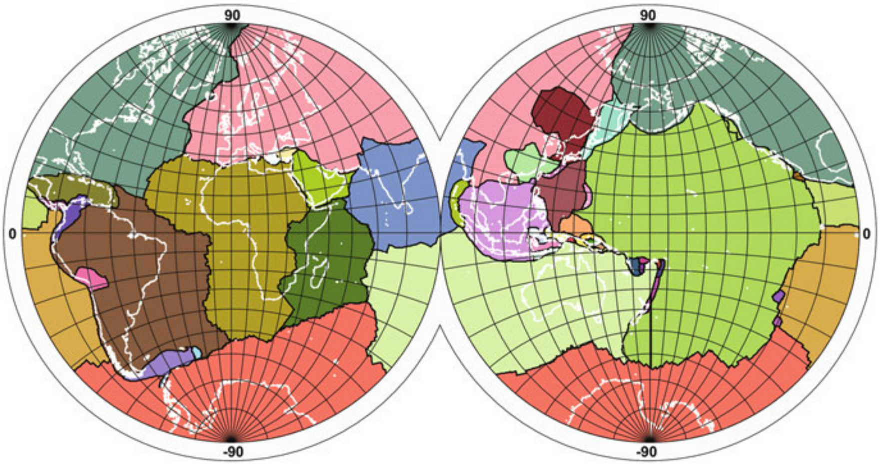

Fig. 8.10 Current tectonic plate boundaries on Earth, indicated with *dark outlines*. Continent outlines are indicated in *white*. Large plates occupy cratonic areas and oceanic basins, while (mostly) convergent zones host locally smaller plates, with very complex geometrical relationships between them. Source: Bird (2007) and later integration by Ahlenius, Nordpil, Bird, on GitHub

measuring absolute time differences at many measurement sites; without such information, plate tectonics could not be demonstrated. Such methods are currently not available on other planets; independent of the existence or absence of clues coming from other types of observations and methods. Determining whether plate tectonics ever operated on other planets is beyond current technological feasibility.

Before the onset of plate tectonics, a global tectonic system called *sagduction* tectonics or drip tectonics is thought to have prevailed, which produced the granite-greenstone belts that characterize Archean tectonics. *Granite-greenstone belt* is a descriptive term that accounts for the green colour of metamorphosed mafic to ultramafic rocks associated with granites and gneiss; attempts to understand their tectonic context has led to the definition of *ultra-hot orogens*, as opposed to the *collisional orogens* that are a product of plate tectonics. Structural evidence shows that crustal deformation of ultra-hot orogens resulted from sagduction, a vertical tectonics mechanism in which denser rocks at the surface sink into lighter rocks. Sagduction may be driven by an inverse density gradient between supracrustal rocks, mainly mafic or ultramafic, and underlying lighter, granite-gneiss rocks; or by lateral shortening and downward extrusion of the supracrustal rocks. This mechanism implies that the radiogenic heat production in the crust was high enough for the crust to be ductile over its whole, or most of its thickness.

8.3.2 The Tectonic Style of One-Plate Planets

Each terrestrial planet has its own tectonic style, resulting from differences in water abundance, internal heat (hence planetary mass), crustal thickness and temperature, and also their accretion history. The number of plates in the terrestrial sense is determined by the number of blocks that are internally rigid and peripherally deformed. Although this looks simple, this criterion asks fundamental questions such as the minimum dimensions of rigid portions of the crust that can be reasonably treated as plates, or rather *micro-plates*, and how much their kinematics needs to be different from the neighboring plates. Even today, determining the number of plates still critically depends on the availability of data. With the increasing number of kinematic observations worldwide, the number of lithospheric plates on Earth has increased from 6 in the first model of plate motions proposed by Le Pichon in 1968 to up to 50 (including many micro-plates) in 2014 (Fig. 8.10).

Perhaps because their geologic evolution has been simpler than that of the Earth, or perhaps because of the difficulties in identifying plates on other planetary bodies, they are usually all considered to be planets with only one plate. Should this view change in the future, and some other planetary bodies would possess a collection of plates, this would not necessarily be evidence of plate tectonics in the terrestrial sense, with accretion, subduction, collision, and strike-slip boundaries. We shall see that the Moon, Mercury, Venus, and Mars can be considered as one-plate planets (Fig. 8.11) but still each having their own style, dominated—as far as it is understood today—by impact cratering (Moon, Fig. 8.12), tidal despinning and secular cooling (Mercury), ultra-hot orogeny (Venus), and mantle plumes (Mars, Fig. 8.13).

8.3.2.1 The Moon

Tectonism on the Moon was (and is) dominated by impact cratering, directly or indirectly (Fig. 8.12). Impact cratering itself (Chap. 7) includes tectonic processes, the style of which depends on the rheology of the target. Impacts that are energetic enough to generate melting have two additional tectonic effects: related magma emplacement before cooling, and then again the creation of stresses once the magma has cooled and loaded the crust.

Infilling of the lunar craters and basins by basalt has various effects. Lava flooding is made possible by magma sheets (dikes and sills), the geometry of which may or may not be affected by the fractures produced by the impact and the gravitational modification of the crater or basin. Dikes and sills are hydrofractures that may produce significant deformation of the surrounding rocks if emplacement is shallow enough. Some narrow grabens that cut the surface of the maria have been interpreted to be a manifestation of such late-stage dikes.

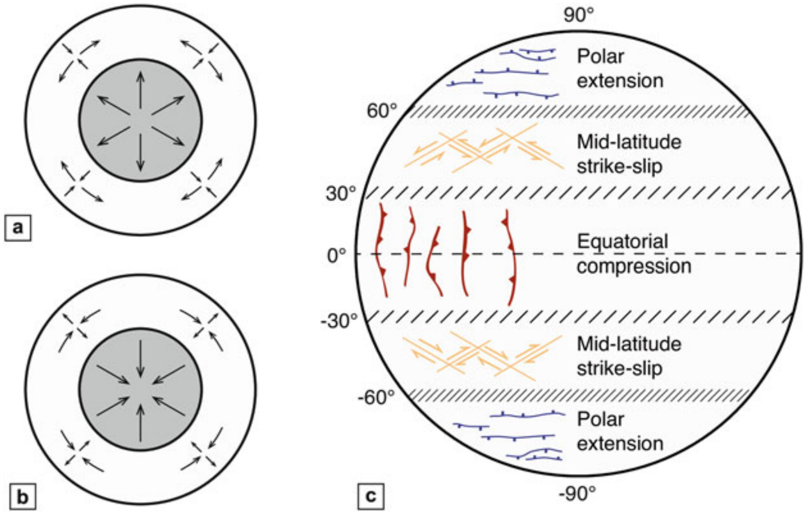

Fig. 8.11 Basic conceptual models of tectonic patterns of one-plate planets. Global radial and concentric stress patterns on one-plate planets due to secular heating and cooling: (**a**) Initial global heating and expansion, with dominant lateral tensional stresses in the crust. (**b**) Late-stage global cooling and contraction. (**c**) Predicted global tectonic pattern of a despinning planet. As the rotation rate decreases, the rotational flattening of the planet decreases and the polar and equatorial radii will increase and decrease, respectively. Correspondingly, different patterns of stress will develop at different latitudes. Source: (**a**, **b**) Redrawn after Solomon (1978). (**c**) Redrawn after Melosh (1977)

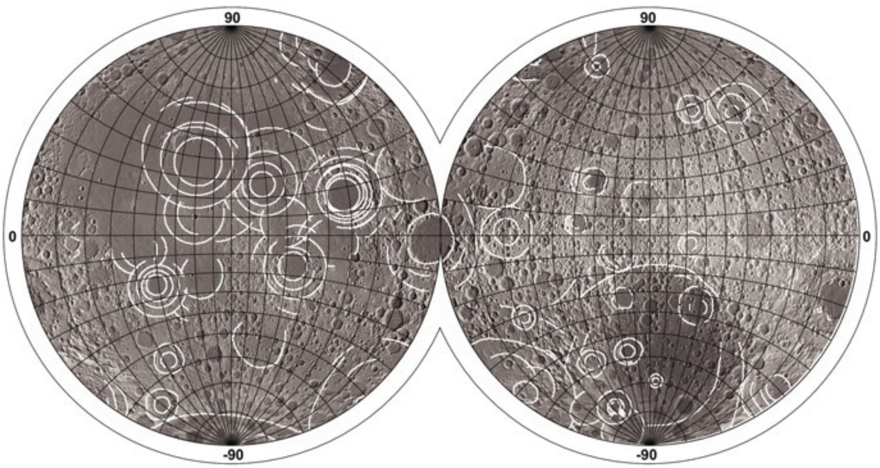

Fig. 8.12 Extensional structures related to large impact basins on the Moon. Contractional structures, mainly within basaltic maria, are not indicated on the figure. Source: redrawn after Geiss and Rossi (2013) based on USGS data from Lucchitta (1978), Scott and McCauley (1977), Stuart-Alexander (1978), Wilhelms (1979), Wilhelms and El-Baz (1977), Wilhelms and McCauley (1971)

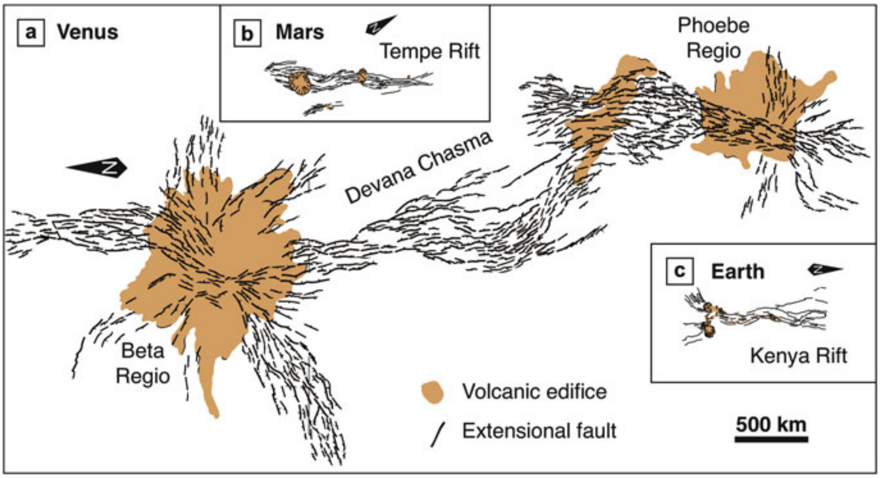

Fig. 8.13 Tectonic sketch map of Devana Chasma, a major rift-like extensional structure on Venus. (**a**) Devana Chasma is comparable in size and structural architecture to (**b**) other extensional systems on Mars as well as (**c**) terrestrial continental rifts such as those in the East African Rift System (e.g., Kenya Rift). Note the large volcanic centres of Beta and Phoebe Regio, which are linked by the rift system. Source: Mapping by P. Kronberg, plus, modified from Hauber and Kronberg (2005)

Lava pool cooling in impact craters and maria has a contractional effect on the surface at the center of the maria, generating wrinkle ridges. The contraction is compensated by extension at grabens that accommodate thermal stress release at the boundary between the cooling lavas inside the maria and the surrounding cool crust.

8.3.2.2 Mercury

One-plate planets may display homogeneous or heterogeneous crustal deformation; Mercury is the one displaying the highest degree of homogeneity. Due to its proximity to the Sun, the crust, which has an estimated average thickness of 35 km $\pm$ 50%, is thought to be dry, preventing any large-scale tectonic motions.

Although the surface of Mercury looks approximately like the surface of the Moon, it differs by the similar, mainly basaltic, composition of the impact structure fill and floors and their host crust exposed around them, as opposed to the difference between lunar basaltic mare and anorthositic highlands, and by the widespread surface deformation by contractional structures. Extensional structures are very rare, though not absent. The abundant contractional features (wrinkle ridges and lobate scarps) do not display an obvious global tectonic fabric, although it was recently shown that compression at mid-latitudes is preferentially E-W oriented. Contraction is sometimes associated with strike-slip deformation. These structures are significantly overprinted only by the Caloris impact basin. There has been little variation in the possible sources of deformation that were investigated since the Mariner 10 flybys (1974–1975), including two global types of events, secular planetary cooling and early tidal despinning (Fig. 8.11).

The effect of secular cooling on the observed global contraction of Mercury has been little questioned, as cooling produces a net decrease of planetary size, necessarily resulting in net surface contraction while tolerating strike-slip faulting. Tidal despinning has been questioned because in the absence of any stress field of other origin active at the same time, the predicted associated deformation also includes extensional faulting. However, when secular contraction and tidal despinning are simultaneously active, the predicted tectonics is fully compressional, with the mid-latitudinal preferential E-W contraction favored, and contraction of random orientation close to the pole is predicted, as observed.

The Caloris region displays additional distinctive tectonic patterns, especially the remarkable graben system of Pantheon Fossae that radiates from the Caloris Basin center and probably formed by surface uplift, for instance following the growth of an underlying intrusion. Based on structural similarity between Pantheon Fossae and structures such as novae on Venus, it was suggested that they formed coeval with dike emplacement at depth.

8.3.2.3 Venus

Despite its similar size and proximity to Earth, which suggests that its bulk composition, inherited from accretion, is also similar at first order, Venus currently lacks plate tectonics and is thought to be a one-plate planet. The high temperature of the crust of Venus, which, in contrast to Mercury, is not due to the proximity to the Sun, but to the high temperature of the atmosphere, does not allow water to be stable within the crust either. This is one of the reasons to explain the absence of plate tectonics. Why there is no water on present-day Venus is not fully understood. Furthermore, the hot surface temperature of $\sim$460 °C was recently shown to cause potential crustal discontinuities to heal, preventing any plate boundary from even initiating.

Volcanic plains occupy more than half of the Venusian landscape. They are interrupted by broad volcanic rises, huge (*crustal*) plateaus, and the Ishtar Terra region. The volcanic plains, and some volcanic rises, include chasmata and coronae structures with diameters ranging from 60 to several hundreds of kilometers (with the exception of Artemis Corona: 2600 km), and a stellate variant called novae. Venusian coronae have no recognized equivalent in the Solar System. Additionally, the volcanic plains display various types of small-scale deformation features (wrinkle ridges, fracture fields, polygons). The volcanic rises include rift zones (Fig. 8.13), and the crustal plateaus display tesserae, a complex superimposition of folds, faults, and lava flows reminiscent of a parquet geometry. Tesserae are, however, not always associated with plateau topography. Ishtar Terra is a huge, unique high area that includes tesserae in the east, the Maxwell Montes orogen in the center, and in the west a high, smooth plateau, Lakshmi Planum, surrounded by compressional belts.

Not all the tectonic features may be due to internal dynamics; the polygonal and other deformation features of the volcanic plains have been shown to be possibly a consequence of climate warming, from a period of moderate greenhouse to the present atmospheric state. Given the difficulty of rigid plate motions on present Venus, the tectonic features listed above are interpreted in terms of dominantly vertical motions, i.e., variants of upwellings and downwellings.

In contrast, coronae are interpreted as ovoid features forming in the crust in response to buoyant diapirs in the mantle. Coronae have a variety of shapes, from domes to collapsed plateaus, and are surrounded by trenches, which can be reproduced by models of diapiric rising and subsequent spreading in response to cooling. Novae are radiating grabens that are not associated with ovoid structures and trenches and correspond to early stages of corona formation. They are associated with pit craters, and other morphologies of likely magmatic origin (sometimes lava flows), indicating that they are the surface expression of underlying dike swarms. As a consequence, many coronae (47%) are also associated with radial grabens, and their mapping can be used to reveal local stress fields.

Five main rifts zones have been identified on Venus, some of which form a large equatorial rift zone system. Some of the Venusian rifts are structurally very similar to terrestrial continental rifts associated with flood basalt provinces such as the East African Rift System. Analysis of rift topography at Devana Chasma (Fig. 8.13), one of the main equatorial rifts, similarly revealed rift flank uplift.

8.3.2.4 Mars

It has been a long debate how many tectonic plates constitute the Martian lithosphere, with terrestrial-type plate tectonics interpretations proposed since the Mariner 9 mission. All these interpretations fall short at global scale, because each defined plate boundary has planet-wide implications on planets of constant radius. For instance, removing a portion of the lithosphere in a subduction zone needs compensation by forming new lithosphere elsewhere, in a spherical puzzle in which each plate boundary kinematics has influence on the kinematics at the other boundaries. One of the most interesting, and perhaps the most fundamental, issue in Martian tectonics is the nature of the so-called *hemispheric dichotomy boundary*, a topographic transition zone that separates the northern plains and the southern highlands. The topographic difference between the lowlands and highlands is $\sim$2 km, and the transition is marked by a mean slope that does not exceed 3–4° over a distance of a few hundred km to 3000 km. The age of the dichotomy is not very well constrained, although it is considered likely that it formed already very early during the pre-Noachian, making it the oldest large-scale morphologic feature of Mars (Chap. 11). There is no evidence of faulting that could have been involved in the formation of the dichotomy. Gravity and topography data inversion has been used to propose crustal cross sections in which the topographic decrease toward north is associated with a rise of the crust-mantle boundary, which suggests that in a descriptive sense, Mars may be considered a two-plate planet. In a kinematic sense, however, it is highly unlikely that the lowlands and highlands once moved with respect to each other along the boundary. Because of the irregular shape of the dichotomy boundary, such movements would have deformed it intensely, which is not observed. Kinematically, Mars is today best considered as a one-plate planet.

The Tharsis volcanic-tectonic bulge (Fig. 8.14) was built on this dichotomy boundary, which it broadly masks. The bulk volume of Tharsis, the largest known

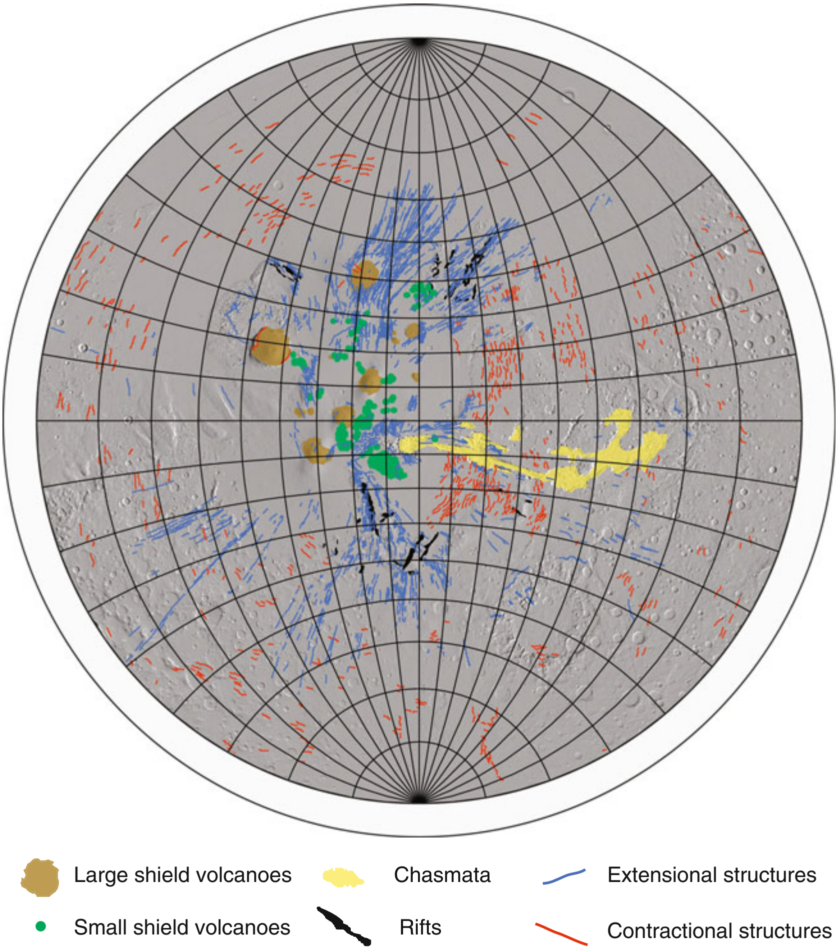

Fig. 8.14 Tectonic sketch map of Tharsis, the largest volcanic province in the Solar System. The huge topographic bulge dominates the western equatorial hemisphere of Mars. It is characterized by very large shield volcanoes (*brownish colors*) and hundreds of smaller volcanic vents (low-shield clusters are shown in *green*), several sets of long and narrow grabens (*thin blue lines*) that radiate outwards from several centers, and a concentric set of wrinkle ridges (*thin red lines*). Volcanic loading of the lithosphere is likely responsible for the concentric tensional stress and the radial compressive stress. A few large and complex extensional features (in *black*) are comparable to terrestrial continental rifts. The ∼3000 km-long trough system of Valles Marineris (*yellow*) is controlled by Tharsis-radial trends and was probably formed by a combination of extension, collapse and erosion. Source: Modified from Hauber and Kronberg (2001)

volcano-tectonic province in the Solar System with a diameter of about 5000 km, formed early in the history of Mars. Its exact formation age is somewhat debated, but its activity, at least locally, continued until relatively recent times (Chap. 11).

Many extensional structures (faults, usually forming grabens, and tension fractures) are radiating from the Tharsis central area, and oriented perpendicular to contractional structures such as wrinkle ridges (Fig. 8.14). These extensional structures may be the surface expression of giant radiating dike swarms, analogous to dike swarms in ancient cratons on Earth.

The intensely deformed zones that predate the Tharsis volcanic flows are organized in *rift zones* that were later reactivated in the context of Tharsis evolution. Rift zones are extensional features that are geometrically and kinematically (but not geodynamically) very similar to terrestrial rifts.

Valles Marineris is a collection of elongated troughs, some interconnected, that extends over 3000 km along strike and encompasses a region 800 km wide. Elongation is parallel to the surrounding Tharsis radial grabens, which are sometimes cut by the trough walls, showing that tectonic extension participated in their formation.

The southern troughs are linear, and triangular-faceted spurs suggest a tectonic origin. The northern troughs, in contrast, are partly or completely closed and have distinctly non-linear plan-view outlines, suggesting dominating (and not fully understood) collapse and erosional processes for their origin. The relative role of tectonic extension vs. collapse in Valles Marineris and related chasmata remains one of the most puzzling enigmas of the geologic evolution of Mars.

8.3.2.5 Icy Satellites and Kuiper Belt Objects

The surfaces of some icy satellites in the outer Solar System display many tectonic features that are not found on the rocky terrestrial planets. Their icy shells are deformed by tidal stresses induced by their parent planet and by other satellites, and their surfaces bear clear records of these stresses (*cryotectonics*). The nature and magnitude of deformation depends on orbital parameters (e.g., eccentricity) and the satellites' obliquity (i.e., the tilt of their rotational axes with respect to their orbital planes) as well as on the thickness of the icy crust, and the rheology of the underlying material. In some cases (e.g., Europa and Enceladus) it seems likely that global subsurface oceans underlie the rigid outer shells, and fractures penetrating the entire crust would perhaps enable the ascent of water to the surface (*cryovolcanism*). As tidal stresses may be considered to be due to external forces, they will not further be discussed in this chapter on endogenic processes.

When the New Horizons spacecraft delivered the first detailed images of the surfaces of Pluto and Charon, one of the biggest surprises was the apparently young age of some surfaces. Obviously, tectonic and perhaps cryovolcanic processes were much more active in the recent geologic history of these Kuiper Belt objects than previously thought. At the time of writing, the spacecraft still transmits data of its flyby, and only preliminary results are available. One of the early findings is that convection in a layer of nitrogen ice forms characteristic polygons with a diameter of 10–40 km. The lack of impact craters in this region, informally called Sputnik

Planum, implies continuous and relatively quick resurfacing. These results add to the emerging view that some small and icy bodies in the Outer Solar System are geologically dynamic bodies (Chap. 13).

8.4 Magmatism and Volcanism: Driving Forces

8.4.1 Igneous Volcanism

On Earth, three main geodynamic settings associated with volcanism may be generally distinguished: volcanism along convergent and divergent plate boundaries, and intra-plate volcanism. Magmatic activity concentrates along plate boundaries and so do volcanoes. The circum-Pacific *Ring of Fire* is the most active and widely known example today. It is mainly located at convergent plate boundaries and associated subduction zones. Even more important in terms of flux and volume, new crust is continuously formed at plate boundaries along the mid-oceanic ridges in a process called sea floor-spreading.

On divergent plate margins, lithospheric plates move apart. As a consequence, the overburden load of the thinned crust above the mantle decreases, hence mantle material can move upward to fill the space between the moving plates. The adiabatic ascent and the decrease in lithospheric pressure enable the generation of adiabatic melts without the need of additional heat—a process known as decompression melting of mantle material. This partial melting generates basaltic magma with low silica contents. An important consequence of volcanism at or along ocean ridges is *hydrothermal activity* where cold seawater is heated and enriched in various chemical constituents.

At convergent plate margins, an oceanic plate is subducted underneath either a continental or oceanic plate, and the subducted oceanic crust descends into the mantle. The delivery of wet crust, carrying water-rich sediments with it, into the hot mantle causes partial melting of the mantle above the subducted plate, as water reduces the mantle liquidus temperature. Melt generation, ascent, and subsequent fractionation and differentiation towards the surface commonly produce high-silica magmas with substantial volatile contents (mainly H_2O, CO_2). Such magmas would often erupt explosively.

Based on our current understanding, all other terrestrial planets are one-plate planets characterized by so-called stagnant lid regimes (Chap. 10). As a result, volcanism, i.e. magma generation and extrusion is typically explained by heat transfer through *mantle plumes* or large-scale mantle upwellings.

Mantle plumes are thought to be hot, buoyant diapiric upwellings of solid mantle material that rise through a planet's mantle in a convective mushroom-like manner, carrying heat from the depth, most often assumed to be the core-mantle boundary, to the surface over a long period of time. Once the top of the plume is close to the surface, the drop in pressure may allow the temperature to cross the solidus causing the generation of a large volume of melt in the upper mantle. These melts may then intrude into the crust or erupt onto the surface to form flood basalt lavas. The locations of such volcanic activity are called *hot spots*. The melting associated with mantle plumes generally produces large volumes of melts periods of time; however, there is a decrease in magma volumes over time as the plume may laterally spread out in the mantle beneath the crust.

On Venus and Mars, magmatism may have been largely associated with mantle plumes, and the style of volcanic activity may reflect this *plume tectonic regime* and would be analogous to hot spot volcanism on Earth, enabling the formation of large-scale volcanic regions with giant volcanic edifices, such as in the Tharsis and Elysium provinces on Mars.

Under certain circumstances, volcanism may dominate the heat transport on planetary bodies from their interior to the surface (Chap. 10). If this is the case, a body enters a *heat-pipe mode* where heat is transferred to the surface by magma ascending throughout the lithosphere in narrow channels. This may be the case for Jupiter's moon Io, the most volcanically active body in the Solar System. The mechanism that produces the internal heat of Io is widely agreed to be dissipation of tidal energy from interaction with Jupiter, induced by the orbital resonance 4:2:1 between Io, Europa, and Ganymede.

The mechanism of cooling by heat-pipes was also proposed for early Earth before plate tectonics developed, as well as for other terrestrial planets early in their evolution.

8.4.2 Non-igneous Volcanism

Volcanism, however, is not restricted to the production of silicate magmas (igneous volcanism). Several icy moons in the Outer Solar System (mainly moons of gas giants and dwarf planets) display evidence for heat transport by cold fluids and volatiles. Such phenomena are referred to as *cryovolcanism*, because they occur in environments with extremely low temperatures. During cryovolcanism, explosive eruptions or effusions are triggered by fluid water and/or aqueous solutions of several other chemical components such as ammonia, methane, nitrogen, hydrocarbons etc. Active cryovolcanism was observed by spacecrafts on Saturn's mid-sized icy satellite Enceladus and on Neptune's largest moon Triton (Chap. 12).

Another type of non-igneous volcanism is *mud volcanism*. During this process a mixture of liquid water and sediment is mobilized and extruded at the surface where it can form a *mud volcano*, which may only be a few metres high. This mobilization of soft sediment can be driven by gases (e.g., CH_4, CO_2 and others); however, mud volcanoes may also be formed by non-volcanic processes, especially in sedimentary basins. For example, kilometre-scale mud volcanoes in Azerbaijan are associated with hydrocarbon deposits from which methane is released into overlying sediments. Additionally, mud volcanoes may be formed also by tectonic processes such as overpressurization in compressional settings. Mud volcanism is a common phenomenon on Earth, and was also suggested to explain the origin of several types of small pitted cones on Mars.

8.5 Magmatic Activity

8.5.1 Composition

The primary product of mantle melting is magma with basaltic composition, and therefore the majority of volcanic rocks on Earth are basalts. Basalts are generated in all volcanic settings, but they are especially dominant at mid-oceanic ridges and intraplate hotspots (flood basalt provinces). More evolved magmas are typically associated with convergent plate margins, where new granitic crust is being produced. On other planets, the dominance of basaltic igneous rocks is even more pronounced. Chemical and mineralogical information, morphologic characteristics, and available samples all point to basaltic compositions for most volcanic rocks on Venus, Mars, Mercury, and the Moon. Nevertheless, there are exceptions. It is known that there are felsic rocks on the Moon, e.g., granites and rhyolites, although they are rare and volumetrically insignificant. It has been speculated that the tessera terrains on Venus may have evolved compositions and may be the equivalent to terrestrial granitic continents. Recently, observations by orbiters and rovers hinted at evolved compositions on Mars, ranging from quartz diorite and granodiorite to trachy-andesite and trachyte. These rocks would represent evidence for low-density crust on early Mars (Fig. 8.15). The discovery of the low-pressure, high-temperature

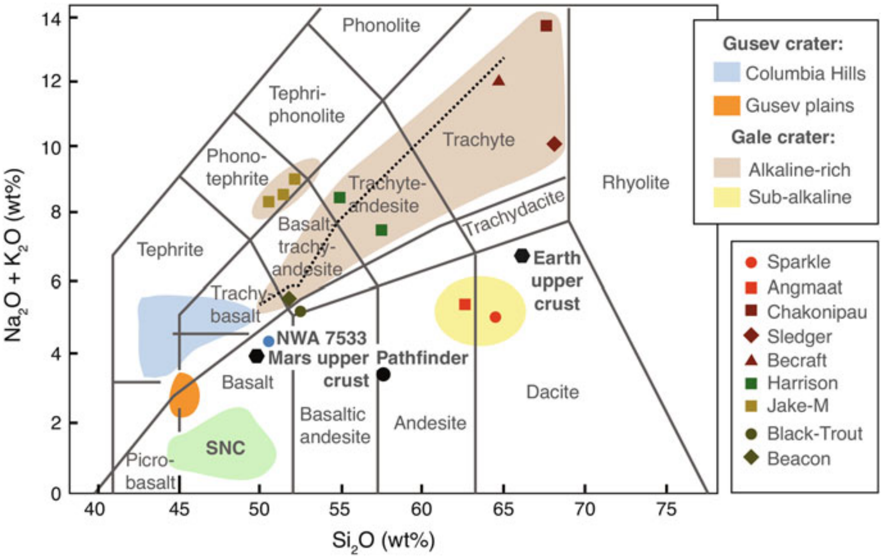

Fig. 8.15 TAS (Total Alkali versus Silica) diagram showing the range of observed crustal compositions of Mars. The *symbols* in the legend on the lower right correspond to samples analysed by the MSL rover, Curiosity, in Gale crater. Such trachy-andesitic, trachytic, and dacitic compositions may represent an early Martian crust that may be compared to continental-type compositions on Earth. The SNC green field includes all Martian meteorites except the Noachian breccia NW7533. Source: Modified from Sautter et al. (2015)

($\sim$870 °C) SiO_2 mineral, tridymite, by the MSL Curiosity rover is the first in situ evidence for silicic volcanism and provides further evidence for evolved magmas on Mars.

More exotic compositions such as sulfur flows were thought to be possible on Io, but the observed extremely high eruption temperatures ($\sim$1,500 K) on Io basically rule out this possibility, as sulfur volcanism would not generate temperatures above roughly 700 K. It is now generally thought that lavas on Io consist mainly of superheated basalts (emplaced after rapid ascent from a deep, high-pressure source) or, alternatively, of cooled ultramafic komatiites. Nevertheless, the actual composition is unknown and remains one of the biggest critical questions in the study of Io.

8.5.2 Plutonism/Intrusions

Melts, once formed, are typically less dense than their parent materials (water is an exception), hence they have a tendency to rise towards the surface. The behaviour of magma (melt plus crystals) near the surface is then dependent on the contrast between the density of the magma and the surrounding rocks. If the magma is less dense than the surrounding rocks, it may reach the surface and erupt. However, if the magma's density is similar to that of the surrounding rocks, the level of neutral buoyancy is reached, causing the magma to stall where it typically forms magma reservoir(s). Over time such reservoirs may fully crystallise and form intrusive igneous bodies of various sizes (e.g., batholiths, laccoliths, etc.).

On Earth, the relative proportion of intrusive versus extrusive rocks (the I/E ratio) is typically estimated to range from 5 up to 40, although this depends on the tectonic environment and the properties of the crust through which magma is rising. Specifically, new crust produced at mid-ocean ridges has a higher than usual percentage of intrusive rocks. Nevertheless, it is also expected that the I/E ratio increases with increasing crustal thickness: the thicker crust acts as a filter, therefore a smaller fraction of magma reaches the surface. For instance, for Io or the Moon, which have very thick and very cold lithospheres, it is expected that the I/E ratio would be possibly much higher than on Earth. As an example, it is commonly believed that the concentration of mare basalts on the lunar nearside is a function of the different crustal thicknesses, with the crust being thicker on the lunar farside. Large magma intrusions have an important effect on the magnetization of the crust, because they may thermally demagnetize the shallow crust beneath volcanoes through heat conduction and associated hydrothermal activity.

Some structural features on Mercury's surface such as floor-fractured craters and radiating grabens may suggest intrusive activity, but overall the evidence is limited. This may be due to the dominance of contractional deformation, which may tend to prohibit the emplacement of shallow intrusions such as dikes and sills. On the Moon, intrusive bodies are thought to exist in association with grabens, and in floor-fractured craters, interpreted to be a surface manifestation of sills beneath them.

On both bodies, Mercury and the Moon, dike intrusions might originate from large depths, and the existence of shallow magma reservoirs does not seem to be common. On Venus, many surface features are likely associated with intrusions; hence, many large and deep magma reservoirs or diapiric upwellings together with both shallow and deep intrusions are expected to be present. On the icy satellites, non-silicate intrusive bodies were proposed as an explanation for linear ridges on Jupiter's rocky-icy moon, Europa.

8.5.3 Effusive Volcanism

Evidence of effusive volcanism is common on all terrestrial planets and Io. For example, the majority of volcanic terrains on the near side of the Moon is associated with extensive lava plains infilling lowlands and large impact craters. They are known as lunar *maria* and they resemble lunar equivalents of terrestrial *flood basalts*. The mare basalts represent by far the majority of volcanic rocks on the Moon, yet they occupy only about 17% of the lunar surface, and have a volume of $10^7\,km^3$, or 1% of the crustal volume. They were mainly formed early in lunar history, and the peak flux occurred during the Imbrian period about 3.85–3.2 Gyr ago. Nevertheless, younger mare flows have been identified, and recent observations suggest that endogenic activity on the Moon persisted until much later and may be as young as one billion years, or even less. On Mercury, 27% of the planet is covered by flat areas termed as *smooth plains* which have many similarities with lunar maria and are mostly located in the northernmost hemisphere and around some large impact basins such as Caloris, Beethoven, and Rembrandt. These plains do not display any unambiguous volcanic landforms, but the morphologic and structural characteristics are consistent with a volcanic origin. The analysis of crater populations that are either buried by or superposed on the smooth northern plains suggests that the volcanic plains have been emplaced over a geologically relatively short time interval of perhaps less than 100 Myr. They represent a minimum volume of 4×10^6 to $10^7\,km^3$ and are comparable to terrestrial Large Igneous Provinces (LIP). It seems likely that they formed in a single, voluminous volcanic event associated with extensive partial melting of Mercury's mantle at around the same time (between $\sim$3.7 and 2.5 Ga), similar to the major episode of mare volcanism on the Moon.

On Venus, about 70% of the surface is covered by volcanic lava plains, indicating that effusive volcanism played a major role for the formation of the currently visible surface. The Venusian surface has a relatively small amount of impact craters suggesting a young age, therefore, an intense period of effusive activity had to occur relatively recently, between $\sim$1 and 0.3 Ga. The more or less random distribution of impact craters on Venus also suggests that volcanic resurfacing was a truly global process. It is not known what kind of endogenic process might have been responsible for such an intense pulse of new crust formation. On Mars, flood basalts are also common; they occur as lava plains surrounding the volcanic centres of Tharsis and Elysium, but are globally widespread as so-called ridged plains. They seem to be possibly related to a period of intense volcanic resurfacing in the Early Hesperian, about 3.6 Gyr ago.

Evidence for localized volcanic centres of various sizes is also common on Mars and Venus. For example, the entire surface of Venus is speckled with large central shield volcanoes, such as Maat Mons or Sif Mons, which are associated with lava flows emanating from their vents. The surface of Mars is characterized by large effusive volcanoes such as Olympus Mons (Fig. 8.7), but also by small effusive volcanic centres, e.g., large portions of Tharsis and Elysium are covered by young extensive lava flows (Fig. 8.9).

8.5.4 Explosive Volcanism

When ascending magma is fragmented into small volcanic particles, called *pyroclasts*, an *explosive eruption* occurs. Basically, there are two possibilities how explosive eruptions originate and how magma might be fragmented. One can be considered as a dry process, in which the eruption is driven solely by volcanic gases originally dissolved in the melt. It occurs when magma ascends relatively quickly from depth and is accompanied by rapid decompression. On Earth, the most abundant volcanic landforms produced by this type of moderately violent explosive activity are *scoria cones*, small conical volcanoes composed of unconsolidated scoria material. Evidence for such magma degassing is known from the Moon, Mercury, Io, and from Mars. Venus has a very dense atmosphere which hinders rapid degassing, and it is debated whether explosive volcanism occurs on Venus or not.

If volcanic gases are trapped in magma by some form of blockage of the conduit, e.g., by a viscous lava dome, pressure inside the volcano builds up and can eventually blast the blockage away. The sudden release of pressure causes rapid gas expansion, intense magma fragmentation, and hence creation of fine-grained volcanic ash. On Earth, such violent explosive eruptions are common for composite volcanoes rising above subduction zones. On Mars, some of the old highland volcanoes (e.g., Hadriacus, Tyrrhenus, and Apollinaris Montes) may have been formed by a similar process.

The second possibility how magma can be fragmented are *wet* (*phreatomagmatic*) eruptions which occur when magma of all types is mixed with an external source of volatiles, e.g., on Earth, seawater or groundwater. The basic principle of this interaction is rapid heat transport from magma to the volatile component, leading to volatile vapourisation, steam expansion and pressure build-up, and explosive magma fragmentation. On Earth, small volcanoes such as *tuff rings* and *tuff cones* are the most abundant volcanic landforms associated with this process. Which specific type of deposit is formed by a phreatomagmatic eruption depends, among other factors, on the amount of water available to drive the magma fragmentation (water-magma ratio). Phreatomagmatic eruptions should have occurred on Mars and Io, where the contact between lava and underlying frozen water or SO_2 may have led to explosive volatilization and to the formation of plumes.

8.5.5 Environmental Effects

The specific planetary environment plays a crucial role determining the type of volcanic activity and hence the shape of volcanic landforms. The environment consists of a complex array of factors, where each of them contributes differently on various aspects of volcanic activity. For example, the gravitational acceleration affects the depths at which magma reservoirs are likely to be formed, but it may also affect the geometry of magma chambers and feeder dikes. With decreasing gravitational acceleration, the sizes of these bodies need to increase to enable buoyant ascent through the country rock. Additionally, gravity affects the distance to which volcanic pyroclastic particles may be ejected in an explosive eruption, thus the ejected particles travel farther from the vent with decreasing gravitational acceleration (Fig. 8.16).

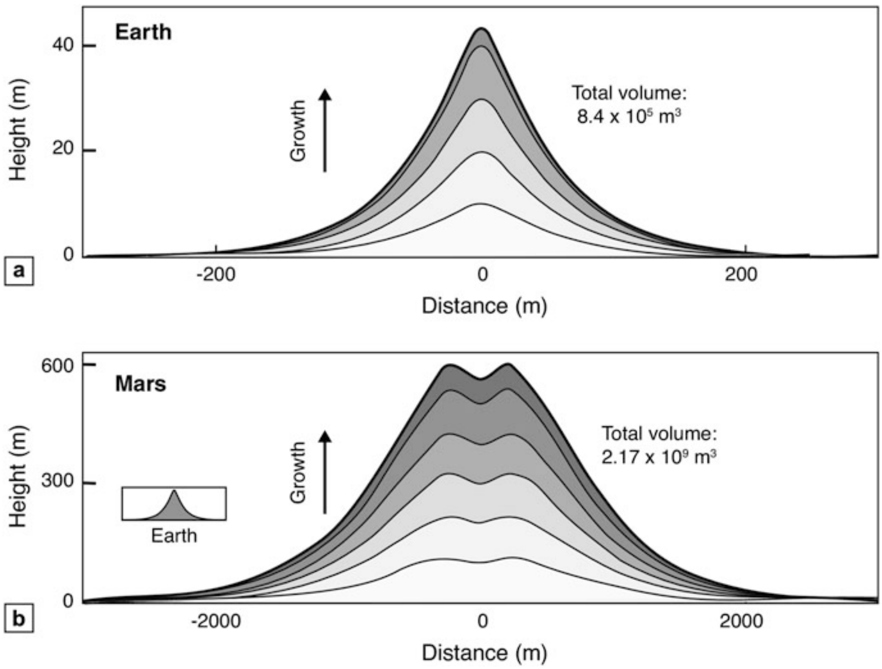

Fig. 8.16 Example of effects of environmental conditions on volcanic eruptions. (**a**) Evolution of a terrestrial scoria cones until the angle of repose (30°) is reached. *Increasing darkness of the fill* indicates the gradual growth of the cone, the *thick solid line* shows profile when angle of repose is reached, corresponding to a height of ~40 m and a volume of about 8.4×10^5 m^3. (**b**) The equivalent height of a cone on Mars is ~600 m, and the angle of repose is reached when the volume of deposited material is about 2.17×10^9 m^3, four orders of magnitude larger than on Earth. Note the dramatically different cone sizes on Earth and Mars when the angle of repose is first reached, a consequence of the much lower gravitational acceleration and atmospheric density on Mars, which enables ejected particles travelling much farther from the vent and thus covering a much larger area. Source: modified from Brož et al. (2014)

Another factor is the atmospheric surface pressure, or the absence thereof. On Earth, subaerial eruptions occur at an ambient pressure of ~0.1 MPa, but this value is quite different on other terrestrial bodies. For example, on Venus, the pressure may be up to ~9.2 MPa (92 bar), preventing volatiles to be easily released from magma. At the other extreme end of ambient pressure conditions, near vacuum exists on Mercury, the Moon, and Io. The absence of an atmosphere capable of inhibiting the exsolution of volcanic gases means that there is a significant drop between the pressures in the ascending magma column and the surrounding environment. Therefore, volcanic gases expand violently and rapidly when magma reaches the surface, causing ejection of magma clasts with high speeds. The ejection of volcanic particles is then characterized by the formation of an umbrella-shaped plume, as such environments do not support the rise of buoyant convective ash plumes as it is common on Earth. Actually, active extraterrestrial volcanism was first detected when a ballistic plume was seen against the dark space in Voyager images of Io (Fig. 8.1).

8.5.6 Outgassing

Outgassing is an especially important phenomenon shortly after the planetary accretion and core formation, when the surfaces of terrestrial planets may be covered by a global magma ocean (Chap. 11), from which large quantities of gases may have been released. This first period of intense outgassing produces a primary atmosphere, which over time is eroded and removed by solar wind and cosmic radiation. When in a later stage of planetary evolution magma reaches the surface, large amounts of dissolved gases, mainly water vapour, carbon dioxide, and sulphur dioxide can be released into the environment, a process called *outgassing*. Together with the impact delivery and the reactivation of subsurface reservoirs, volcanic outgassing is the main source of volatiles for the formation of a secondary atmosphere. Volcanism has a profound impact on the evolution of planetary atmospheres, and potentially influenced the early climate on Earth, Venus, and Mars. On shorter timescales, volcanism is historically known to have significant effects on the climate. On Titan, the atmosphere probably originated from ammonia which was released on the surface by massive volcanic activity, and was later converted to nitrogen by ultraviolet photolysis.

8.6 Volcanic Characterization of Solar System Bodies

8.6.1 The Moon

Compared to Venus and Mars, the inventory of volcanic landforms on Mercury and the Moon appears relatively limited, both in terms of diversity and abundance. On both bodies, extensive plains of basaltic lavas are the dominant volcanic features. On the Moon, these plains (*maria*) consist of mare basalts, which are well visible from the Earth as large dark areas. Mare basalts form very leveled (equipotential) surfaces and lavas could flow over very long distances ($\sim$1200 km), which is thought to be a result of their very low viscosities. Vent structures are rarely observed, and typically the lava flows cannot be traced back to their sources. A particular type of landform associated with mare volcanism are *sinuous rilles*, meandering channels that often start at a circular or elongate depression and gradually disappear into the lava plains of the maria. The lengths can range from a few kilometers to more than 300 km. There is a consensus among many researchers that the channels are formed by a combination of mechanical and thermal erosion of the underlying substrate.

Regional dark mantle deposits, mostly associated with uplands adjacent to younger mare regions, are interpreted to be pyroclastic materials, probably dispersed over tens to hundreds of kilometers by Hawaiian-style fire fountaining driven by continuous gas exsolution. Pyroclastic glass formed by lava fountaining of gas-rich, low viscosity and Ti-Fe-rich basaltic magmas is volumetrically negligible but scientifically important as it is derived from melts generally unaffected by crystal fractionation. Such glasses represent the best samples of the lunar mantle. In comparison to mare basalts, the non-mare volcanic features are relatively insignificant. Most notably, some obviously volcanic domes can reach diameters of 20 km and heights of $\sim$1000 m. Prominent examples are the Gruithuisen domes, whose shapes suggest that they consist of viscous lava, and the Mairan domes, which might be explosive in nature.

8.6.2 Mercury

Large volcanic plains related to impact basins and in high northern latitudes do not display broad, rifted rises, constructional landforms (e.g., shield volcanoes), or individual linear, leveed lava flows. They were probably emplaced by flood-lava eruptions, and are not the products of narrow, leveed flows sourced from small dikes and more limited-volume surface eruptions. The volume of the northern volcanic plains is estimated to be at least $4 \times 10^6\,km^3$ to $10^7\,km^3$, consistent with high-volume, high effusion-rate eruptions. Such volume is comparable to that of terrestrial LIPs (Columbia River Basalts: $1.3 \times 10^6\,km^3$; Deccan Traps: $8.2 \times 10^6\,km^3$; Greater Ontong Java igneous province: $\sim 7 \times 10^7\,km^3$), and indeed the northern volcanic plains on Mercury are considered to be a planetary analogon to terrestrial LIPs.

Mercury was thought to be volatile-poor, hence no signs of explosive volcanic activity were expected prior to recent spacecraft exploration with high-resolution cameras. However, dozens of deposits have now been identified that are morphologically consistent with emplacement by explosive activity.

8.6.3 Venus

As for the tectonic landforms, the larger bodies (Venus and Mars) display a higher abundance, morphological diversity, and age range of volcanic landforms than the smaller bodies, Mercury and the Moon. Again, this is not surprising, as volcanism is linked to the endogenic activity of a planet, which in turn is coupled to its mass and tectonic evolution. Venus displays large volcanic provinces with huge individual edifices (shield volcanoes) that can reach diameters of more than 1000 km, and extensive lava flow fields that can cover plains with a lengths of ~1000 km. Rectilinear to sinuous *channels* within Venusian lava plains can reach extreme lengths of ~6800 km (Nile river: ~6500 km). Lavas with extremely low viscosities are required to build channels of such lengths, and basaltic lavas typically do not have such low viscosities. Other compositions like ultramafic komatiites, high-Ti lunar-type basalts, and carbonatites or sulphur flows may be candidate alternatives. Venus also shows many small individual volcanic cones or shields, which form volcanic shield fields. The high atmospheric surface pressure is thought to prevent rapid and violent magma degassing, analogous to the high water pressure on Earth's ocean floors. Some researchers have suggested that submarine volcanic landforms may indeed be useful analogues for the study of volcanism on Venus. Although some landforms have been tentatively interpreted as products of explosive eruptions, including a pyroclastic flow deposit, no conclusive morphologic evidence of explosive eruptions has been observed on Venus. On the other hand, there are several very large surface features on Venus that do not have terrestrial analogues. They appear to have a volcano-tectonic origin and have been classified as magmatic centers. In many cases, they have axisymmetric planform geometry and are characterized more by structural features indicating surface deformation than by magmatic features like volcanic edifices or lava flows. This class of features includes *coronae*, *arachnoids*, and radial fracture centers, or *novae*. Coronae can have diameters of up to 2000 km and are characterized by a concentric ring of tectonic fractures that surround other fractures and volcanic landforms such as domes and flows. Coronae as well as other large magmatic centers are thought to be related to mantle plumes (upwelling), but downwelling models have also been proposed. Recent thermal observations from orbit were interpreted as indicative of active volcanism, but due to the lack of modern high-resolution radar imaging it is not possible to search for associated landforms.

The distribution of volcanoes suggests that magmatism and internal dynamics on Venus are driven by large-scale mantle convection and mantle plumes, based on similarities with the distribution pattern of hot-spot volcanism on Earth. Indeed, Venus may be considered an observable analogue for the Archean Earth before the onset of plate tectonics (Chap. 11).

8.6.4 Mars

The surface of Mars shows numerous volcanic landforms of diverse morphology, which were produced by both effusive and explosive eruptions. The best-known examples are huge basaltic shield volcanoes. The largest of them, Olympus Mons, has a basal diameter of ~600 km, a height of more than 20 km, and is probably the largest volcano in the Solar System. These edifices share many physiological characteristics with large terrestrial shields: very low flank slopes (~5°), and numerous large individual lava flows that can be channel-fed or tube-fed. It seems that these volcanoes were built over a very long time, perhaps billions of years. Huge calderas formed in several stages, and show that large shallow magma chambers must have existed beneath their surface (Fig. 8.7). The youngest lava flows on their surfaces are only a few tens of million years old; some may even be as young as a few million years. Lava flows on Mars can reach extraordinary lengths of several hundred kilometres, indicating high volumetric flow rates for extended periods of time. Lava flows of this length are rarely, if ever, observed on the Earth, and it is debated whether their formation on Mars requires some extraordinary conditions. The rheology of Martian lava flows, as determined from their morphology and morphometry in remote sensing data, suggests basaltic lavas with low viscosities.

Hundreds of much smaller shield volcanoes are distributed in clusters within the Tharsis and Elysium provinces (Fig. 8.14). Their diameters range from several to tens of kilometers, their heights reach a few hundred meters, and the corresponding flank slopes are extremely shallow (~0.5°).

Since methane in Mars' atmosphere was first reported on the basis of orbital and Earth-based observations back in 2004, and later confirmed by MSL rover measurements, researchers tried to explain how it is currently supplied to the atmosphere where its survival time is short. Since a biologic origin seems unlikely, possible sources may involve geologic processes such as serpentinization. On Earth, methane is a common product of mud volcanism (sometimes also called sedimentary volcanism). Therefore, several types of pitted cones have been interpreted as mud volcanoes in the wake of the possible methane detections. As the shape of mud volcanoes can resemble that of igneous volcanoes (e.g., monogenetic scoria cones or tuff cones), interpretations can be ambiguous, and in several cases both origins (igneous and sedimentary) seem possible.

8.6.5 Io

As a consequence of tidal heating by Jupiter, Io experiences a very high surface heat flow and a volcanic activity that surpasses that of the Earth. Volcanic landforms are abundant and can be classified into a few groups: Volcanoes, pyroclastic materials, and mountains that are probably no active volcanoes, but have a volcano-tectonic origin. Calderas on Io tend to be very large (average diameter ~40 km, maximum diameter 200 km), hence this may hint at larger sizes of magma chambers than on Earth. Some very large lava flows on Io are considered to be analogues to terrestrial flood basalt fields.

8.6.6 Icy Bodies

Although unambiguous morphological evidence for cryovolcanism is rare on icy bodies in the outer Solar System, it has been demonstrated that it is an active process on Enceladus and Triton and was possibly active on Europa, Titan and Kuiper Belt objects such as Pluto and Charon. One definition of cryovolcanism is the extrusion of liquid water and gases that would be frozen solid at the ambient surface temperatures of the respective body. The energy to drive cryovolcanism is thought to come from tidal forces. Cryovolcanism is very important for the emerging scientific discipline of astrobiology, as it is a form of exchange between subsurface oceans that are thought to underlie the icy shells of some satellites, and the surface and space. As such, it delivers plumes to space that could potentially contain volatile species and organic compounds that may be biosignatures. Sampling such cryovolcanic plumes by spacecraft would then be an efficient method to analyse such materials, whereas the penetration of icy planetary shells by, e.g., melting probes may technologically not be feasible for some time to come.

Take-Home Messages

Most, but not all volcanic landforms on Venus, Mars, Mercury, and the Moon have morphological analogues on the Earth, and seem to be predominantly basaltic in composition.

The larger terrestrial planets do have a greater diversity in volcanic morphology than the smaller ones, and the presence of water/ice on Mars makes it the most diverse besides the Earth.

Differences in morphometry, shape, and other properties (e.g., particle size distribution), may be explained by the different environmental conditions, e.g., the atmospheric pressure and the gravitational acceleration.

There are only a few tectonic landforms on the terrestrial planets (e.g., the coronae on Venus) that do not have direct counterparts on the Earth.

It is expected that the continuing exploration of the Solar System with spatially higher-resolving instruments in the future (e.g., better radar imaging of Venus) will enable identifying new and smaller-scale endogenic landforms.

Of special interest are small rocky-icy bodies such as Europa, Enceladus, Titan, Triton, and Kuiper Belt objects such as Pluto and Charon, as they seem to be much more geologically active than expected.

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