10.09 Planetary Tectonics and Volcanism: The Inner Solar System

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10.09.1 Introduction

The Earth is but one of the many planetary bodies in the solar system (in this chapter, a ‘planetary body’ encompasses any solid body orbiting the Sun that is larger than a few hundred kilometers in diameter) and therefore represents only one datum. It is essential to examine geologic processes on other planetary bodies to gain the most complete understanding. If a model works well on Earth, a vital test for the validity of that model is to determine if it works equally well when applied to places in the solar system with different ambient conditions (e.g., gravity, surface pressure, and atmospheric composition). Thus, the investigation of volcanic and tectonic processes on other planetary bodies provides insight into those same processes operating on Earth.

In the last 20 years, terabytes of data have poured down to Earth from spacecraft orbiting Venus, Mars, the Moon, Jupiter, and even a few asteroids. Information gleaned from these missions has revealed extraterrestrial processes that in some cases are quite familiar and, in others, remain poorly understood. In this chapter, the current understanding of volcanic and tectonic processes operating in the inner solar system (Mercury, Venus, the Earth/Moon, and Mars) are summarized. A single chapter cannot possibly cover all the information available – the entire outer solar system is avoided here – and therefore, the reader is encouraged to pursue additional readings (see, e.g., Watters and Schultz, 2010).

10.09.2 Planetary Volcanism and Tectonics

10.09.2.1 Tectonics

Both volcanic and tectonic processes are controlled largely by lithospheric and asthenospheric behaviors on Earth; other planetary bodies large enough to experience differentiation (larger than a few hundred kilometers in radius) should also have (or have had in the past) similar mechanical layering. Earth’s tectonics are dominated by the lateral movement of rigid lithospheric plates over the weaker asthenosphere. This style of tectonism – plate tectonics – is unique to Earth and localizes tectonic activity primarily to plate boundaries. The other terrestrial planets (see Table 1) display movement and deformation of a single, continuous lithospheric shell. On the Moon, tectonics are locally controlled by large (>10^2 km radius) impact basins. Mars’ tectonics are regionally controlled by both large impact basins and volcanic provinces (such as the Tharsis region). Large impact basins have tectonic signatures on Mercury, but there are also more global patterns of tectonism that are unique among the terrestrial planets. Venusian tectonism is expressed almost globally across the Venusian plains. Locally, quasicircular tectonic terrains called ‘coronae’ (up to thousands of kilometers across) and rift valleys called ‘chasmata’ reveal lithospheric extension and compression.

The icy satellites of Jupiter and Saturn also clearly display tectonic features, but there are fewer data available to constrain the mechanical layering inside those planetary bodies. Observations to date support that most of these icy satellites are also ‘one-plate planets’ whose tectonics are controlled mostly by large impact basins and tidal forces, with volcanic processes exhibiting local dominance.

10.09.2.2 Planetary Lithospheres

Tectonics are the movements of the lithosphere, so to understand planetary tectonics, it is first necessary to learn about the properties of these extraterrestrial lithospheres and the limitations of that knowledge (see Chapter 10.05, and references...
Table 1  Physical parameters of the terrestrial planets

<table>
<thead>
<tr>
<th>Planet</th>
<th>Equatorial radius (km)</th>
<th>Bulk density (kg m⁻³)</th>
<th>Surface gravity (m s⁻²)</th>
<th>$T_E$</th>
<th>Atmospheric pressure (Pa)</th>
</tr>
</thead>
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<tr>
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<td>2439</td>
<td>5430</td>
<td>2.78</td>
<td>172</td>
<td>0</td>
</tr>
<tr>
<td>Venus</td>
<td>6051</td>
<td>5250</td>
<td>8.60</td>
<td>22</td>
<td>$9.12 \times 10^6$</td>
</tr>
<tr>
<td>Earth</td>
<td>6378</td>
<td>5520</td>
<td>9.78</td>
<td>34</td>
<td>$1.01 \times 10^8$</td>
</tr>
<tr>
<td>Moon</td>
<td>1738</td>
<td>3340</td>
<td>1.62</td>
<td>103</td>
<td>0</td>
</tr>
<tr>
<td>Mars</td>
<td>3393</td>
<td>3950</td>
<td>3.72</td>
<td>126</td>
<td>600</td>
</tr>
</tbody>
</table>

$^a$From Melosh (2011), Table 4.1.
$^b$Value given for sea level (Earth) or mean planetary radius.

equation [1] can give quite different values from those calculated using other methods. For example, Zuber et al. (2010) estimated $T_E$ for Mercury to be on the order of 30–60 km, based on topographic measurements of contractional features. Smith et al. (2012) used orbital data collected from the MESSENGER (MErcury Surface, Space ENvironment, GEochemistry, and Ranging) spacecraft to estimate Mercury’s $T_E$ to be only 70–90 km. These variations reflect a combination of data resolution and the complexities (such as lithospheres composed of layered materials with different rheologies) inherent in the parameters controlling $T_E$.

Lithospheric or crustal deformation is responsible for the tectonic features observed on other planetary bodies. The crust, a portion of the lithosphere, is distinguished from the lithosphere by composition: in differentiated bodies, the crust is the outermost layer, distinguished by having the lowest density. On Earth, the crust is the outermost portion of the lithosphere, and it is likely that similar arrangements exist on other terrestrial planets. The large density contrast between the crust and mantle makes it somewhat simpler to model the crustal thickness than the lithospheric thickness.

Modeling the thickness of the crust and lithosphere requires detailed measurements of the surface topography and gravity. These data sets currently exist for Mars and Venus; MESSENGER data from Mercury and GRAIL (Gravity Recovery and Interior Laboratory) data for the Moon (see Table 2) are improving our modeling capabilities for those planetary bodies. Mars’ crust displays a strong dichotomy: in the southern highlands, the crust is ~60 km thick – about twice as thick as in the northern lowlands (Neumann et al., 2004). On Venus, the spatial resolution of the gravity and topographic data provide fewer model constraints; recent work (James et al., 2013) suggests that the mean crustal thickness on Venus is 8–25 km. Data from GRAIL indicate a lunar crustal thickness between 34 and 43 km (Wieczorek et al., 2013). Acknowledging that the available data are relatively poor quality for the southern hemisphere of Mercury, Smith et al. (2012) calculated a crustal thickness ranging from ~20 to 80 km, with a mean of 50 km.

10.09.2.3 Mare-Type Wrinkle Ridges

Mare-type wrinkle ridges (or, simply, ‘wrinkle ridges’) appear on every terrestrial planet (Watters, 1988), but are relatively rare on Earth (Figure 1). These features were first described on the lunar surface, where they are linear to arcuate in planform and have a superficial resemblance to folds (or wrinkles) in a rug – hence their name. Complete wrinkle ridges consist of a broad topographic arch topped by a narrow sinuous ridge (Golombek et al., 2001; Watters, 1988); locally, the sinuous ridge may meander off of the broad arch. They are asymmetrical in a topographic profile drawn perpendicular to their long axis, and the sense of asymmetry may change along ridge length. Wrinkle ridges are now recognized to be blind (not intersecting the surface) thrust faults (Golombek et al., 2001;
through layered materials, with surface folding, although the precise nature of the faulting remains unclear (Golombek et al., 2001; Schultz, 2000; Watters, 2004; Zuber, 1995). The size and spacing of parallel wrinkle ridges can be used to determine the total thickness of the deformed material (e.g., Watters, 1993; Zuber, 1995): large spacing (10–10^2 km) reflects a greater thickness of deformed material than does small spacing (10s of kilometers) (Montesi and Zuber, 2003a,b; Watters, 2004; Zuber and Aist, 1990).

Figure 1  Mare-type wrinkle ridges (black arrows) are the most common tectonic feature in the inner solar system. Apollo 17 metric frame showing southern Mare Serenitatis (Apollo Image AS17-M-0451 (NASA/JSC/Arizona State University)). Black arrows point to mare-type wrinkle ridges; note broad arch and superposed sinuous ridge. White arrows point to troughs, interpreted to be graben. Image centered at 19.9° N, 24.1° E.

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10.09.3  The Moon
10.09.3.1  Lunar Tectonism
Lunar tectonics are dominated by large (>10^2 km across) impact basins (Watters and Johnson, 2010; Wilhelms, 1987). Wrinkle ridges and linear or arcuate troughs are commonly found within or adjacent to impact basins (Watters and Johnson, 2010). Wrinkle ridges and lobate scarps are compressional features; linear or arcuate troughs are generally interpreted to be graben and therefore extensional. Lobate scarps are more broadly distributed across the lunar surface (Watters and Johnson, 2010) and are interpreted to be thrust faults (Binder, 1982; Watters et al., 2010).

Lunar impact basins on the nearside are filled with various amounts of basalts; these lowlands are referred to as maria (mare = singular). Wrinkle ridges are found within the maria and are typically both radial and concentric to the basin center (Maxwell et al., 1975; Watters and Johnson, 2010). Wilhelms (1987) pointed to wrinkle ridge rings within large lunar basins as marking the location of buried basin ring complexes, suggesting that premare topography may play an important role in the spatial distribution of lunar wrinkle ridges. Alternatively, lithospheric loading by basaltic lava may have caused basin subsidence, resulting in concentric and radial compressional stresses (Freed et al., 2001; Maxwell et al., 1975; Solomon and Head, 1979).

Linear or arcuate troughs (graben) are also associated with large lunar impact basins and their maria. They are interpreted to be graben because of their flat floors, inwardly dipping walls, and symmetrical topographic cross sections (Figure 1): debate remains about the precise nature of these graben: they may be the expression of near-surface dikes (Head and Wilson, 1993) or have been formed by localized basin stresses (Golombek, 1979). These graben are typically located entirely within maria deposits (and therefore are likely caused by basin-controlled extension) but locally may extend for a few kilometers into the adjacent highlands. No graben are observed on the lunar farside (Watters and Johnson, 2010) until recently: the Lunar Reconnaissance Orbiter Camera (LROC) revealed small-scale (locally as shallow as ~1 m) graben in the farside highlands and mare basalts (Watters et al., 2012a). These graben may be as young as 50 My (Watters et al., 2012a).

A unique class of impact crater, floor-fractured craters display a complex, spider-web-like network of graben on their otherwise flat floors (Figure 2; Wichman and Schultz, 1995; Wilhelms, 1987). Alphonsus crater (2.1° E, 13.5° S) is a type example. The current consensus is that these types of graben networks are formed by localized uplift of the crater floor, although the specific cause of the uplift remains unclear. Results of numerical modeling (Dombard and Gillis, 2001) suggest that viscous relaxation of the uplift remains unclear.

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explanation for floor fracturing of larger impact craters (Wichman and Schultz, 1995).

The spatial correlation of wrinkle ridges and linear or arcuate graben with lunar maria strongly suggests a causal (and therefore temporal) relationship. Certainly, many of the largest lunar impact basins on the nearside are associated with mass concentrations (‘mascons’) that could also be related to deformation of basin-weakened lithosphere (Solomon and Head, 1979, 1980). Crosscutting relations and impact crater size–frequency distributions suggest that both wrinkle ridges and graben formed shortly after the maria were deposited. The detailed examination of maria deposits cut by lunar graben, combined with impact crater size–frequency distributions, suggests that no maria younger than 3.6 Ga (/C6 0.2 Ga) are crosscut by linear or arcuate graben (Lucchitta and Watkins, 1978). In contrast, wrinkle ridges are observed to deform the oldest (~4.0 Ga) and the youngest (1.2 Ga) maria (Hiesinger et al., 2003).

Lobate scarps (interpreted to be thrust faults (Schultz, 1976)) have been estimated to be even younger than the youngest wrinkle ridges, based on their relatively fresh appearance (Figure 3) and impact crater size–frequency distributions (Binder and Gunga, 1985). It has been proposed that some lunar lobate scarps may even currently be active (Watters et al., 2010, 2012a).

10.09.3.2 Lunar Volcanism

Lunar volcanism is concentrated on the lunar nearside within impact basins. These maria cover ~6.3 × 10⁶ km² (Head, 1976) with a volume of ~5.0 × 10⁶ km³ (Budney and Lucey, 1998; Hiesinger et al., 2002; Horz, 1978). The Apollo astronauts returned over 350 kg of lunar material (NASA, 2007), and it is primarily from these samples (as well as rare meteorites from the Moon that have struck the Earth) that we know the composition of lunar lavas. By using these lunar samples as ‘ground truth’ for comparison with satellite data, the composition of the lunar surface is fairly well constrained.

The lunar maria are ‘basalts,’ but these are unlike any basalts observed on Earth, and there is a range in lunar basalt compositions (see Table 3; Shearer et al., 2006). Although both lunar and terrestrial basalts share a relatively low SiO₂ content (<54 wt%), the abundances of the remaining oxides belie this similarity. Lunar basalts have more iron and less aluminum than typical terrestrial basalts, giving them a high melting temperature (~1400 °C) and a low viscosity (<1 Pa s at liquidus temperature) (Murase and McBRhney, 1970). Furthermore, H₂O and CO₂ are the most abundant volcanic volatiles on Earth (e.g., Schmincke, 2004); in the anhydrous and reducing environment of the Moon, CO is the most probable volatile (Wilson and Head, 1981). In addition to these

<table>
<thead>
<tr>
<th>Oxide</th>
<th>The Moon: Apollo 11, hi-Ti, low-K (#10020)²</th>
<th>Nakhla meteorite (Mars)²</th>
<th>Venus² Venera 14 Landing site</th>
<th>Earth basalt²</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
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<td>48.6</td>
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</tr>
<tr>
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<td>1.68</td>
<td>18.4</td>
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</tr>
<tr>
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<td>9.0</td>
</tr>
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<tr>
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<td>12.1</td>
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<tr>
<td>CaO/Al₂O₃</td>
<td>1.14</td>
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</tbody>
</table>

²BVSP (Basaltic Volcanism Study Project) (1981).
⁴Kargel et al. (1993).
⁵Best and Christiansen (2001).
compositional differences, the lunar volcanic environment has a lower gravity, no atmosphere, and a surface temperature that ranges from 126 to 373 K (see Wilson and Head, 1981). It should not be surprising, therefore, that lunar volcanism displays different morphologies than we see for basaltic volcanism on Earth.

Lunar basalts are concentrated in large impact basins on the nearside, although they also exist as lava ‘ponds’ (<2000 km²) (Yingst and Head, 1997). Maria that are obscured by superposed, higher albedo material (such as impact crater ejecta) are termed ‘cryptomaria’ (Antonenko et al., 1994). The stratigraphic position of cryptomaria suggests that it represents some of the oldest volcanism on the Moon. However, this same stratigraphic position makes it difficult to identify and study.

The low viscosity of lunar basalts makes the preservation of lava flow margins rare, although some fantastically large (>1200 km long) examples can be seen in the Imbrium basin (Schaber, 1973; Schaber et al., 1976). Furthermore, low-viscosity lavas are unlikely to pile up around the vent to create tall volcanic constructs. Finally, low-viscosity lavas are likely to flood low-lying terrain, potentially burying their own vents. Thus, although lunar lavas comprise 17% of the lunar nearside, vents and edifices are relatively rare and tend to be concentrated in a few geographic locations: the Marius Hills region, the Aristarchus plateau region, and the Grünthiesen domes.

The Marius Hills region (Figure 4) is characterized by mounds and hills that have been interpreted to be volcanic domes and cones (Weitz and Head, 1999). Domes have flank slopes of 2° and cones are steeper; both have basal diameters <25 km. All currently available data are consistent with these hills being basalt (rather than a more evolved, silica-rich, viscous lava). The analysis of recently acquired topographic data from the Lunar Orbiter Laser Altimeter led Spudis (2011) to suggest that the entire Marius Hills complex is a large (330 km across and 3.2 km tall) shield volcano peppered with smaller parasitic vents. Lawrence et al. (2013) stated that the Marius Hills represent eruptions of mare materials and the morphological variability was caused by a range of eruption and emplacement conditions.

Surrounding the Marius Hills is a high spatial density of sinuous rilles (Figure 4). Sinuous rilles are long (up to hundreds of kilometers) sinuous troughs that maintain a relatively constant width (generally <5 km) along their length. These troughs are generally u-shaped in cross section (as opposed to v-shaped) and some of them originate in circular or irregular depressions (Greeley, 1971; Schubert et al., 1970). These rilles formed via lava, although there remains debate as to whether they are erosional (lava downcutting into the preexisting surface by melting and assimilating the country rock) or constructive (e.g., lava tubes whose roofs have collapsed over time or drained lava channels) features (e.g., Williams et al., 2000 and references therein). There are hundreds of sinuous rilles on the lunar surface, and it is unlikely that all of them formed in precisely the same way: it is probable that some combination of constructive and erosional processes is responsible for their formation (Roberts, 2013).

The low viscosity of lunar basalts and lack of water would predict an abundance of effusive volcanism (i.e., lava flows) rather than explosive volcanism. However, astronaut Jack Schmidt returned a sample of orange soil from Apollo 17; analyses revealed that the soil contained microscopic droplets of orange volcanic glass. Subsequently, green glass droplets were identified in the Apollo 15 samples. These droplets formed by explosive eruptions of lunar basalts (Meyer, 2005); in the absence of atmospheric drag, the magma fragmented by bubbles (probably of CO) popping on the lunar surface formed microscopic beads. Dark mantle deposits (DMDs) are easily observed on the floor of Alphonsus crater (Figure 3), and the spatial association of these DMDs with small craters and the floor fractures (interpreted to be graben) indicates that the DMDs are pyroclastic deposits (Weitz et al., 1998). DMDs are now recognized across the lunar maria (Weitz et al., 1998), representing explosive lunar eruptions.

The timing of lunar volcanism is determined through a combination of (1) radiometric dating of Apollo-returned samples, (2) careful geologic mapping to determine maria stratigraphy, and (3) impact crater size–frequency distributions. Radiometric dating of Apollo-returned samples gives ages of 3.6–3.9 Ga (Apollo 11 and Apollo 17 high-titanium basalts) to 3.16–3.4 Ga (Apollo 12 and 15 low-titanium basalts) (Nyquist and Shih, 1992; Snyder et al., 2000). Mapping and impact crater size–frequency distributions suggest that the youngest maria patches on the lunar nearside may be only 1.0–1.2 Ga (Hiesinger et al., 2003; Schultz and Spudis, 1983). Lava production on the Moon appears to have peaked 3.5–3.8 Ga (Shearer et al., 2006), indicating that the Moon is a volcanically ‘old’ planet. However, there is evidence that Ina, a D-shaped crater about a kilometer across (18.6° N, 5.3° E), may still be actively degassing (Schultz et al., 2006) although the amount and type of volatiles remain unknown.

10.09.4 Mars
10.09.4.1 Martian Tectonics

Mars, like the Moon, is a one-plate planet (Figure 5), but can be divided into three distinct physiographic, topographic, and geologic provinces: (1) the northern lowlands, (2) the southern highlands, and (3) the Tharsis region (see Golombek and
Phillips, 2010 and references therein). The northern hemisphere is topographically low and relatively smooth, whereas the southern hemisphere is rugged with impact craters and is topographically higher (McCauley et al., 1972); this is commonly referred to as the ‘crustal dichotomy.’ Locally, the dichotomy boundary is marked by a scarp that may be as tall as 5.5 km (Aharonson et al., 2001; Watters et al., 2007); elsewhere, volcanics and erosional processes have modified (or buried) the nature of the boundary. There is evidence for some movement along this scarp in the form of compressional (lobate) scarps, extensional features such as graben, and broad topographic signatures of lithospheric flexure (Watters, 2003a,b; Watters et al., 2007). The similar ages for terrains north and south of the dichotomy boundary (e.g., Frey, 2006; Watters et al., 2007) argue against a plate-tectonic genesis for the dichotomy boundary. The origin of this dichotomy remains a mystery (e.g., Citron and Zhong, 2012), but it appeared early (>3.6 Ga) in Mars’ history (Frey, 2006; Frey et al., 2002; Hartmann, 2005; Hartmann and Neukum, 2001; Nimmo et al., 2008; Watters et al., 2007).

Although ancient seafloor spreading has been proposed as a possible origin for the crustal dichotomy (Sleep, 1994), there is no morphological or chronological evidence for plate tectonics at the dichotomy boundary (e.g., Frey, 2006; Pruis and Tanaka, 1995). However, there is an interesting magnetic signature in the oldest portions of the southern highlands (Acuna et al., 2001). The alternating bands of strong/weak magnetic fields in the southern highlands are reminiscent of the alternating magnetic bands found in the ocean crust around Earth’s mid-ocean ridges and require that Mars once generated its own active global magnetic field. The location of the magnetic pattern requires that Mars’ dynamo stopped working within a few hundred million years of planetary formation (Acuna et al., 2001; Hartmann and Neukum, 2001), leaving behind no geologic signature of rifting associated with the magnetic stripes.

The Tharsis region is the largest volcanic province on Mars, and its tectonic influence spans over half of the planet. Volcanic products (primarily lava flows) constructed a pile ~5000 km across and 9 km tall during the first billion or so years of Mars’ history (Carr and Head, 2010; Phillips et al., 2001); currently, the volcanic pile is approximately 10 km high. The enormous mass of this volcanic pile deformed the martian lithosphere, creating a circumferential trough, an antipodal high, and gravity anomalies (Phillips et al., 2001). Circumferential wrinkle ridges surrounding the Tharsis region (indicating compression) are particularly visible in Lunae Planum (Watters, 1988, 1993; Zuber and Aist, 1990), where their regular spacing (~60 km; Zuber, 1993) suggests that the thickness of the deformed layer is also ~60 km (a ‘thick-skinned’ interpretation). Alternatively, the wrinkle ridges may be the result of deformation of only the volcanic plains (a few kilometers thick – a ‘thin-skinned’ interpretation) (Watters, 2004). Most striking is the largest canyon in the solar system, Vallis Marineris (4000 km long, up to 7 km deep, and 200 km wide) (Figure 5). The parallel, inward-dipping walls and flat floor of this canyon are consistent with Vallis Marineris being a large rift valley system. Geophysical modeling results show that the distribution and orientation of most tectonic features around the Tharsis region can be explained by flexural loading of the lithosphere, using present-day values for gravity and topography (Golombek and Phillips, 2010). These results, combined with the timing of the deformation, indicate that the martian lithosphere beneath the Tharsis region has remained unchanged since about 3.6 Ga. Recent modeling by Beuthe et al. (2012) suggests that the elastic lithosphere thickness varies throughout the Tharsis region, being thickest beneath Olympus Mons and thinner elsewhere.

However, the thickness of Mars’ elastic lithosphere is not everywhere the same nor was it likely the same at every stage in Mars’ history. Published elastic lithosphere thickness estimates range from a mere 2 km thick beneath portions of Ascræus Mons in the most recent period of Mars’ history to possibly more than 300 km thick beneath Apollinaris Mons during the same time (see Golombek and Phillips, 2010, and references therein).

Mars displays a staggering variety of tectonic features: wrinkle ridges (thrust faults) and linear to arcuate troughs (graben) are abundant, but strike-slip faults are rare (Andrews-Hanna et al.,

![Figure 5](image_url)
Martian tectonic features are concentrated around volcanic provinces and large impact basins. As technology and image resolution improve, smaller and smaller tectonic features can be observed, but the same general patterns are still evident.

Martian wrinkle ridges are morphologically similar to their lunar counterparts (Figure 6) (Schultz and Watters, 2001; Watters, 1993, 2003a). They are most commonly found deforming plains materials that were emplaced around 3.2–3.6 Ga (Hartmann and Neukum, 2001; Hartmann, 2005). Near the Tharsis region (e.g., in Lunae Planum), wrinkle ridges are circumferential to the approximate center of the Tharsis region and their topography steps downward away from the Tharsis region, consistent with a thrust fault origin (Watters, 1993). Away from the Tharsis region, in Hesperia Planum, wrinkle ridges form roughly orthogonal, crosscutting networks that are oriented radial and circumferential to Hellas basin. Goudy et al. (2005) determined that the radially oriented ridges are younger than the circumferential ridges, indicating a change in the regional stress field over time. Hesperia Planum also reveals ridge rings (Figure 6) that are interpreted to form over buried impact craters (e.g., Watters, 1993). Hesperia Planum deposits are <3 km thick (Ivanov et al., 2005) so it is unlikely these wrinkle ridges deform the entire lithosphere.

Similar to the Moon, Mars also displays lobate scarps that are interpreted to be lithospheric scale thrust faults (Schultz and Tanaka, 1994). They are larger than wrinkle ridges (both longer and taller), and shortening is estimated to be one to two orders of magnitude greater along lobate scarps than along wrinkle ridges (Watters, 1993).

Grabens on Mars range from a few meters across to kilometers across. The interpretation of high-resolution (<3 m per pixel) images shows evidence for erosion and deposition within narrow graben, suggesting that the primary morphology could be obscured (Golombek and Phillips, 2010). The subsurface character of these narrow graben remains unclear. Bounding faults that dip inward at ~60° would intersect at some depth beneath the surface that may represent a mechanical discontinuity. Alternatively, one fault could terminate against a deeper master fault (Schultz et al., 2007). It has been proposed that most martian graben are the expression of near-surface dikes (e.g., Ernst et al., 2001; Mege and Masson, 1996; Scott and Wilson, 2002; Wilson and Head, 2002). In contrast, pit crater chains on Mars – which appear to be genetically related to some narrow graben (Wyrick et al., 2004) – are formed primarily by dilational normal faulting.

**10.09.4.2 Mars’ Volcanism**

To date, all available data indicate that volcanism on Mars is mafic and probably basaltic. Evidence from remote sensing data, volcanic morphology, Mars’ meteorites, and in situ analyses of Martian rocks all support basaltic volcanism (Table 3). Remote sensing detection of possible evolved (silica-rich) lavas in Nili Patera is best explained by being hydrothermal deposits rather than lavas or igneous intrusions (Skok et al., 2010). Mars has a thin (<600 Pa at mean planetary radius, or mpr) atmosphere (roughly 0.6% of Earth’s) that is 95% CO2 (NASA, 2010). Although liquid water is not currently stable on the surface of Mars, there is abundant morphological (e.g., Carr, 1996) and geochemical (e.g., Morris et al., 2010) evidence that liquid water existed on the surface in the past. Water ice is found on Mars today, being most abundant in the polar regions (Bibring et al., 2004; Clark and McCord, 1982) and in the subsurface (Schorghofer and Forget, 2012). Thus, when investigating volcanism on Mars, it is important to look for the evidence of interaction between magma (lava) and water (ice).

There are three main volcanic provinces on Mars: (1) the Tharsis region, which boasts both the tallest (Olympus Mons) and widest (Alba Mons) volcanoes in the solar system; (2) the Elysium region; and (3) the circum-Hellas region, which contains some of the oldest volcanoes on Mars. In addition, Mars appears abundantly covered in ‘ridged plains,’ which are interpreted to be thinly layered basalt flows (Greeley and Guest, 1987; Scott and Tanaka, 1986) that were largely emplaced around the same time (~3.2 Ga). Effusive eruptions appear to have been more common than explosive eruptions, although it is possible that the explosive eruptions have not been fully identified yet (e.g., Kerber et al., 2011).

Both the Tharsis and Elysium regions are dominated by shield volcanoes. The Tharsis region contains four large shield volcanoes: Olympus, Ascreaus, Arsia, and Pavonis Montes, as well as a low shield volcano (Albus Mons) and numerous smaller shield volcanoes (Figure 5). ‘Shield’ volcanoes are so
named because they are characterized by shallowly sloping flanks (4°–11°) that, in profile, look like a Viking shield resting on its handle. The shields within the Tharsis region are morphologically similar to terrestrial shield volcanoes (such as Mauna Loa volcano, Hawaii), complete with summit caldera complexes and volcano rift zones, but are orders of magnitude larger than terrestrial volcanoes. This size difference is likely because Mars does not have plate tectonics, and so, the lithosphere is never moved away from the underlying melt source. The Elysium region contains Elysium Mons and a number of smaller shields. By analogy with terrestrial basaltic shield volcanoes, it is assumed that these Martian shields were fed by an underlying mantle plume (on Earth, a ‘hot spot’).

In contrast, the volcanoes of the circum-Hellas region have flank slopes even shallower than shield volcanoes (<2°), and their flanks are dissected by radial channels. (Note that prior to 2010, these volcanoes are referred to in the literature as ‘paterae’; after 2010, ‘patera’ refers to the summit caldera complexes of these low-lying volcanoes and the volcanic constructs are called ‘montes’ ([IUGG].) As a group, the volcanoes around Hellas are also called ‘hIGHLAND patera’ (Greeley and Spudis, 1981) because they are located in the southern highlands. The low flank slopes and the easily eroded nature of the shield material have led to the suggestion that these volcanoes experienced explosive eruptions and that the shields are composed primarily of pyroclastic materials (e.g., Crown and Greeley, 1993; Greeley and Crown, 1990). Modeling results suggest that the observed distribution of shield materials could be generated by either explosive decompression of mafic gases or magma–water–ice interactions as rising magma encountered ground ice or groundwater (Greeley and Crown, 1990; Gregg and Farley, 2006). Impact crater size–frequency distributions and stratigraphy indicate that the explosive activity is ancient (>3.5 Ga) (Williams et al., 2007, 2008). Younger lava flows (as young as ~1.5 Ga) were subsequently erupted from both Hadriacus and Tyrhensus Montes (Werner, 2009; Williams et al., 2007, 2008). Thus, these volcanoes experienced a change in their volcanic styles with time. Unlike the Tharsis and Elysium regions, the circum-Hellas volcanics may be directly related to the formation of the Hellas basin (e.g., Crown and Greeley, 1993; Greeley and Crown, 1990): it is possible that deep fractures generated by the impact event, combined with mantle upwelling in response to lithospheric thinning, may have created preferred pathways for magma to reach the surface.

Volcanic plains are widespread on Mars (Greeley and Guest, 1987; Scott and Tanaka, 1986) and are interpreted to be composed of low-viscosity basalts (cf. Greeley et al., 2005). Head et al. (2002) proposed that in addition to the visible plains material, there are volcanic plains underlying most (if not all) of the northern lowlands. The type example of these volcanic plains on Mars is Hesperia Planum (Figure 6), characterized by a flat surface lacking obvious vents and lava channels/valleys and deformed by wrinkle ridges. Morphologically, volcanic plains are most similar to terrestrial flood basalts. Globally, these volcanic plains are ancient, emplaced prior to about 3.5 Ga (Werner, 2009).

Impact crater size–frequency distributions strongly support that most volcanism on Mars is ancient: the major volcanoes were close to their present size by 3.6 Ga (Werner, 2009). Determining whether volcanism on Mars remains ‘active’ depends entirely on the definition of ‘active.’ Terrestrial volcanoes are considered to be active if they have erupted in the past 10 000 years, based on the easily observed evidence left by retreating glaciers at that time. There is no similar standard for Mars. However, lava flows that are as young as 1 Ma have been identified (Hauber et al., 2011) through mapping and impact crater size–frequency distributions.

Evidence for lava–water interactions on Mars remains sparse and elusive, although the planetary conditions would favor these interactions in the past (Head and Wilson, 2002). Lava flows with thermal fracture patterns called ‘columnar jointing’ have been observed in impact crater walls (Milazzo et al., 2009); on Earth, the formation of columnar jointing is enhanced by the presence of meteoric water (Grossenbacher and McDuffie, 1995). Rootless cones on Earth form when basaltic lava flows over water-saturated ground, and these have been suggested to exist on specific Martian lava flows (Fagents et al., 2002). Flank slopes of small (<50 km basal diameter) Martian shields increase poleward, and Garvin et al. (2000) suggested that magma interaction with groundwater or ground ice was responsible for the poleward morphological shift. Allen (1979) and Hodges and Moore (1994) used photointerpretation of Viking Orbiter images to identify constructs on Mars that are similar to those formed by subglacial basaltic eruptions on Earth. Gregg et al. (2007) suggested that fluid lavas within Gusev crater (Greeley et al., 2005) ponded against an ice-rich deposit that has subsequently been removed. Chapman (2002) argued that layered mounds within Vallis Marineris were generated by subglacial eruptions. In spite of evidence for glaciers on the flanks of the Tharsis region volcanoes in the past (e.g., Fastook et al., 2008), unequivocal evidence for lava–ice interactions there remains elusive.

### 10.09.5 Venus

Venus is almost the same size and bulk density as Earth (Table 1). Notably, the surface of Venus has an average temperature of ~450 °C and a pressure of 9.6 × 10^5 Pa at mean planetary radius; the atmosphere is mostly CO2. There is no evidence for water vapor in the atmosphere (e.g., Donahue and Russell, 1997) and liquid water is unstable at Venus’ surface; it is likely that the Venusian lithosphere and interior are also anhydrous (Kaula, 1990, 1995; Phillips et al., 1997). The dry nature of Venus’ geologic materials influences their mechanical and physical properties (e.g., Lenardic et al., 1995; Phillips et al., 1997). It may be the lack of water on Venus that is largely responsible for the differences in tectonic and volcanic behavior on Earth and Venus. The limited information available for the composition of the Venusian surface comes from modeling results, measurements made by the Venera landers, and inferences based on volcanic morphology (e.g., Baker et al., 1992, 1997; Gregg and Greeley, 1993; Nimmo and McKenzie, 1998). Overwhelmingly, the evidence points to basaltic volcanism, but there are local exceptions (discussed in the succeeding text).

To understand the scientific context of Venusian tectonism and volcanism, it is necessary to grasp the following concepts: First, there is a lack of impact craters on the surface of Venus...
(only 967 on the entire planet (USGS Astrogeology Branch, 1998)), indicating that the average surface age is ~750 Ma (McKinnon et al., 1997). Second, there remains a controversy about the mechanisms responsible for such a young surface. Venus does not have plate tectonics, and so, that cannot be the resurfacing mechanism (e.g., McGill et al., 2010). Two end-member hypotheses have been proposed to explain Venus’ young surface: (1) catastrophic resurfacing and (2) equilibrium resurfacing (e.g., Basilevsky et al., 1997; Phillips et al., 1992; Strom et al., 1994). In catastrophic resurfacing, the entire surface of Venus is reformed in one fell swoop, possibly through global lithospheric overturn. In equilibrium resurfacing, volcanism and tectonism act throughout space and time to pave a little bit of the surface with lava here and deform a bit of the surface through tectonism there; over time, the entire planet is eventually resurfaced. Regardless, Venus has a geologically young surface compared to Mars, the Moon, and Mercury.

This begs the question: Is Venus ‘active’ volcanologically or tectonically? The data available for the Venusian surface were collected by synthetic aperture radar (SAR) or by Earth-based radar to see through Venus’ optically thick cloud cover. The resolution of the Magellan spacecraft’s SAR instrument was, at best, 75 m per pixel (Wall et al., 1995; Young, 1990); Earth-based radar is on the order of ~2 km per pixel. The Magellan mission collected data from X to Y. During that time, at the best available resolutions, no observable change was detected on the surface. However, the European Space Agency craft Venus Express is currently orbiting Venus and measuring atmospheric properties. Upon arrival at Venus in 2006, the instruments detected a surge in the amount of atmospheric SO2, which then decayed over time (Marcq et al., 2013); this is consistent with a volcanic eruption injecting SO2 into the atmosphere. In the absence of information about changes in surface morphology, however, the hypothesis of active volcanism on Venus cannot be disproven.

### 10.09.5.1 Venus Tectonism

The following tectonic terranes have been identified on Venus: (1) plains, (2) volcanic rises, (3) crustal plateaus and tesserae, (4) coronae, and (5) chasmata (canyons) (Figure 7; Hansen et al., 1997; McGill et al., 2010). In the absence of plate tectonics, most of these features are explained by either mantle upwelling, downwelling, or some combination of the two.

Plains are stratigraphically superposed on other tectonic terranes and so are locally the youngest materials. They are interpreted to be volcanic plains, composed of thinly layered lava flows; this interpretation is supported by images of the Venussian surface taken by the Venera landers (e.g., Garvin et al., 1984; Surkov et al., 1983). Plains are typically deformed by wrinkle ridges, graben, lineaments, and/or polygonal lineaments. Although lineaments are not tectonic landforms per se, they are locally, at least, likely to be expressions of a small (subresolution) structural feature such as a fracture or fault. Wrinkle ridges on Venus, imaged by radar, look different from those on the other planets: Magellan SAR cannot resolve a broad arch and a superposed sinuous ridge. Instead, wrinkle ridges are expressed by a sinuous bright line (Figure 8; note that Magellan images are processed so that bright = high radar backscatter) and are interpreted to represent compressional forces. Commonly, parallel wrinkle ridge sets extend over hundreds to thousands of kilometers, implying a similar stress field over those distances.

Only graben at least 225 m wide can be resolved on the highest-resolution Magellan SAR data; it is inferred that where graben are found parallel with lineaments, the lineaments are also graben but are too small to be completely resolved. Graben and lineaments in radiating swarms, or nova, are abundant on Venus (Head et al., 1992). Based on their morphological similarities to radiating dike swarms on Earth, the Venusian novas are interpreted to be the surface manifestation of radiating dike swarms that did not quite reach the

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**Figure 7** Global Venus topography and SAR. All Magellan images are processed so that bright = high radar backscatter; dark = low radar backscatter. Pinks and whites are topographic highs; blues are lows. (see http://astrogeology.usgs.gov/search/details/Venus/Magellan/RadarProperties/Colorized/Venus_Magellan_C3-MDIR_ClrTopo_Global_Mosaic_6800m/cub).
surface to erupt (Grosfils and Head, 1994). Similarly, most of the graben on Venus are interpreted to be the expression of near-surface volcanic dikes (McGill et al., 2010).

Locally, plains are deformed by polygons. These may be constructed of wrinkle ridges (compression) or graben (extension) or of fractures; individual lineaments are commonly too narrow to resolve completely. The precise origin of these polygons remains unclear.

Ridge belts (formally known as ‘dorsa’) are concentrated zones of compressional deformation within plains materials. They are characterized by a broad (10^2 km) topographic arch, superposed with wrinkle ridges and locally folds (e.g., Banerdt et al., 1997; Zuber, 1990). With one exception (McGill et al., 2010), ridge belts are older than the surrounding plains. Fracture belts are also characterized by a broad topographic arch containing fractures and graben and so are concentrated zones of extensional deformation. They appear to be composed of deformed plains materials.

Ten volcanic rises have been identified to date on Venus, which are 1400–2500 km across, are topographically high (0.5–2.5 km tall), and are spatially associated with some of the largest volcanoes on Venus. Volcanic rises are interpreted to form over mantle plumes, based on topography and gravity data (Phillips and Malin, 1983; Senske et al., 1992); the large compensation depths of some volcanic rises may indicate they are still being actively supported (Smrekar, 1994).

Tesserae and crustal plateaus are both topographically higher than the surrounding materials and are older than the surrounding plains. Crustal plateaus are roughly circular regions 1000–3000 km across that are 0.5–4 km above the adjacent plains, and most contain tessera terrain (Hansen et al., 1999; McGill et al., 2010). Compared with volcanic rises, they have shallow apparent compensation depths, suggesting isostatic compensation (McGill et al., 2010). They formed via mantle upwelling (Grimm and Phillips, 1991; Phillips et al., 1991) or downwelling (Bindschadler and Parmentier, 1990; Lenardic et al., 1995). Ishtar Terra is a unique expression of a crustal plateau on Venus (Hansen et al., 1997): it is a large central plain (containing volcanic calderas) surrounded by mountain belts and Maxwell Montes, the tallest mountain on Venus (Bindschadler et al., 1992; Hansen et al., 1997; Keep and Hansen, 1994; Roberts and Head, 1990; Smrekar and Solomon, 1992).

According to the Gazetteer of Planetary Nomenclature (http://planetarynames.wr.usgs.gov/DescriptorTerms), tesserae are ‘tile-like, polygonal terrain’ but this description does not fully encompass the complicated nature of Venusian tesserae. Tesserae are characterized by multiple intersecting sets of linear to arcuate tectonic features and a high surface roughness relative to the surrounding plains materials (Hansen et al., 1999). Locally, tesserae terrains are the stratigraphically lowest materials, but that does not require that globally all tesserae are the same age (cf. McGill et al., 2010). Given the range of tectonic features and their orientations found in deforming tesserae, it is likely that different patches of tessera represent different episodes of deformation and possibly distinct starting materials (Hansen and Willis, 1996).

Coronae are circular to ovoid features (Basilevsky et al., 1986) 50–260 km across (Glaze et al., 2002) that are typically surrounded by concentric ridges and fractures. Most coronae display volcanic processes: they may contain volcanoes and/or lava flows. They are interpreted to be caused by buoyant mantle material (Hansen, 2003; Smrekar and Stofan, 1997; Stofan et al., 1991) rising from shallower depths than those plumes associated with volcanic rises. Interestingly, coronae are not randomly distributed across the planet, but are concentrated in the Beta–Atla–Themis region (Figure 9; Squyres et al., 1993) and tend to be spatially associated with canyon systems (chasmata).

Chasmata are canyons, interpreted to be tectonic rifts. They are <7 km deep and 10^2–10^4 km long (Solomon et al., 1992). The chasmata contain multiple lineaments that can crosscut and extend beyond the rift; although most lineaments are too small to resolve, locally one can be identified as a trough and so all these lineaments are interpreted to be graben or normal faults (McGill et al., 2010). Individual features, such as impact craters (Figure 10; Solomon et al., 1992) and volcanoes (Stofan et al., 2005), have been rifted apart by chasmata, allowing for extension to be measured: 10–20 km locally (Conners and Suppe, 2001; Rathbun et al., 1999; Solomon et al., 1992).

**10.09.5.2 Volcanism on Venus**

The majority of central-vent volcanoes can be identified as shield volcanoes. Crumpler et al. (1997) had classified shield...
volcanoes as large (>100 km basal diameter), intermediate, or small (<20 km basal diameter). In addition, there are volcanic features on Venus that are unique among the terrestrial planets.

The most abundant volcanic feature on Venus is small (<25 km basal diameter) shields (Figure 11). There are 10^5–10^6 small shields on Venus, and they typically occur in clusters or groups. Individual shields may contain summit pits. Aubele (2009) had identified two different types of shield clusters: shield fields and shield plains. Shield fields are typically 10^2 km across and contain 10^2 shields, whereas shield plains are 10^3 km across and contain 10^5 shields (Aubele, 2009; Miller and Gregg, 2012). Lava flows from shield fields...
...commonly bury surrounding wrinkle ridges, but locally lineaments crosscut shields.

Canali are long (>107 km) sinuous channels found on the Venusian plains and tend to have neither obvious source regions nor sinks. Balta Vallis is the largest example on Venus and is >7000 km long. Along its length, Balta Vallis maintains a consistent width of 3–5 km. The examination of the topography along its length reveals undulations (e.g., Baker et al., 1997) that must have deformed the channel after its emplacement. Long lava flows (103 km long) have been identified on the flanks of large shield volcanoes. For comparison, the longest unconfined lava flow yet identified on Earth is the 160 km long Undara flow in Queensland, Australia (see Stephenson et al., 1998 and references therein). (The Grande Ronde flow of the Columbia River Basalts traveled >1000 km from its source, but the lava was confined in a steep and narrow canyon (Mangan et al., 1986).) Many scientists have questioned whether typical tholeiitic basalts could flow such great distances without solidifying, even on Venus, and have therefore considered the possibility of more exotic lava compositions with low melting temperatures and low viscosities (such as carbonatite and sulfur) (e.g., Gregg and Greeley, 1993; Kargel et al., 1994). The main difficulty with the exotic lava hypothesis is that it is untestable: models of flow behavior confirm that carbonatite or sulfur flows could flow hundreds to thousands of kilometers without solidifying, but petrologic models cannot generate the 107 km3 of exotic lavas required to fill the observed channels on Venus. Without more constraints on the composition of Venus, the likelihood of large volumes of exotic lavas cannot be assessed.

Valley networks observed on Venus must be the result of volcanic processes because liquid water cannot exist at the Venusian surface. Baker et al. (1992, 1997) and Komatsu et al. (2001) classified valley networks based on morphology as simple (e.g., canali), complex, compound, or integrated. Valley networks are found in plains materials and may be the source of lava comprising the plains. Some simple valley networks are morphologically similar to lunar sinuous rilles and probably have similar origins. The complex and compound valley networks, consisting of many tributaries, distributaries, and anastomosing reaches, can only be formed by a long-lived (months to centuries) eruption (Gregg and Greeley, 1993). Integrated networks are morphologically similar to valley systems generated by groundwater sapping processes on Earth; they appear to have been formed by collapse caused by sub-surface removal of material. On Venus, integrated networks may form as lava tube roof collapse or may represent the presence of a low-viscosity, low-melting-temperature lava on Venus, such as carbonatite or sulfur flows (e.g., Baker et al., 1992, 1997; Kargel et al., 1993; Komatsu et al., 2001).

At the other rheological extreme are ‘pancake domes’ (Figure 12), which are steep-sided, flat-topped volcanoes >20 km across and <5 km tall. Some contain a summit pit. They typically occur in groups and are aligned, suggesting that they were emplaced via a dike. Their large aspect ratio (diameter/height) is similar to the aspect ratio of silica-rich (rhyolite) domes on Earth, such as the Owyhee domes in Long Valley, California (Stofan et al., 2000). This observation led to the interpretation that the Venusian domes may also be rhyolitic or at least contain more SiO2 than typical basalts. The Soviet lander Venera 13 returned measurements of the surface composition consistent with alkali basalts, a type of basalt more evolved than tholeite. Pancake domes are located within the Venera 13 landing ellipse (Weitz and Basilevsky, 1993), and it is possible that Venera 13 touched down on top of one of these domes. However, the Venusian domes are 10–100 times more voluminous than terrestrial rhyolite domes, and the petrogenesis of such large quantities of evolved magma in the absence of water or plate tectonics is uncertain. More importantly, detailed analyses of surface roughness and dome topography on Earth and Venus suggest that the Venusian domes are unlike terrestrial silica-rich domes (Stofan et al., 2000).

### 10.09.6 Mercury

Mercury, slightly larger than Earth’s Moon (Table 1), has the largest iron core relative to planetary radius of all the terrestrial planets. It is also the closest to the Sun, and ‘space weathering’ (the interaction of solar ions with Mercury’s surface) affects the physical properties of surface materials in ways that are not yet fully understood (e.g., Robinson et al., 2008; Zurbuchen et al., 2008). It lacks plate tectonics and is a ‘single-plate’ planet like the Moon and Mars. In 1974–75, Mariner 10 made 3 flybys of Mercury and succeeded in mapping almost 45% of the planet at a resolution >1 km per pixel. NASA launched the MESSENGER Surface, Space EnVIRONMENT, GEochemistry, and Ranging (MESSENGER) spacecraft in 2004, and as of this writing, it is in orbit around Mercury (see http://www.nasa.gov/mission_pages/messenger/main/index.html), and almost the entire surface has been imaged at resolutions >250 m per pixel.

#### 10.09.6.1 Mercury Tectonics

Tectonic features on Mercury can be broadly categorized as being either ‘impact basin-related’ or ‘distributed’ (Watters and Nimmo, 2010; Watters et al., 2009a). The Caloris basin (Figure 13) is the largest feature on Mercury and displays one main topographic ring with a diameter of 1550 km. Additional concentric features have been mapped both inside and outside the main ring (Murchie et al., 2008), and these may be additional basin rings. The Caloris basin displays a range of tectonic features that represent ‘basin-related’ tectonics on Mercury (e.g., Watters et al., 2009b).
Tectonic features observed on Mercury include wrinkle ridges, lobate scarps, high-relief ridges, and troughs (interpreted to be graben). Compressional features are far more abundant than extensional ones (Watters and Nimmo, 2010) and are consistent with global contraction (possibly related to the formation of the large iron core). Crosscutting relations suggest that most lobate scarps formed after the emplacement of the smooth plains materials and infilling of Caloris and Rembrandt basins (Solomon et al., 2008; Watters et al., 2005, 2009a) and may be related to local loading (possibly infilling of impact craters with lava).

Lobate scarps are the most common tectonic feature on Mercury (Watters et al., 2009a) and, like those on the Moon and Mars, are interpreted to be thrust faults (Figure 14; Nimmo and Watters, 2004; Watters and Nimmo, 2010). They are considered to be ‘distributed’ and have been observed to crosscut exterior basin fill materials associated with the Caloris basin (cf. Watters et al., 2009b). The oldest materials deformed by lobate scarps are intercrater plains (>4.0 Ga) (Spudis and Guest, 1988; Watters et al., 2009a), and the youngest materials that lobate scarps crosscut are less than ~3.5 Ga old, placing rough constraints on the timing for these features. On Mercury, each lobate scarp represents 1–3 km of shortening (Watters et al., 2009c). The longest lobate scarp yet identified is Enterprise Rupes (centered at 36.5° S, 283.5° W), which is at least 1000 km long (Watters et al., 2009c). Topographic relief across lobate scarps ranges from 0.9 to 2.0 km. Based on the measurements of fault scarp heights and the modeling of thrust fault behaviors, the estimate for Mercury’s \( T_E \) at time of faulting was on the order of 35–40 km (Watters et al., 2002); recent measurements using the MESSENGER Mercury Laser Altimeter (Cavenaugh et al., 2007) allowed Zuber et al. (2010) to estimate values as high as 60 km for Mercury’s \( T_E \).

Wrinkle ridges on Mercury are morphologically similar to those observed on Mars (e.g., Head et al., 2008; Solomon et al., 2008; Strom et al., 1975) (Figure 15); even ridge rings (which form over buried impact crater rims) have been identified (Watters and Nimmo, 2010; Watters et al., 2009a). Wrinkle ridges tend to be found in topographic lows and may therefore be related to basin subsidence (Watters 1988, 1993). Importantly, every type of smooth plains yet identified on Mercury (both those interpreted to be volcanic and those interpreted to be impact-related) is deformed by wrinkle ridges (Watters et al., 2009a).

Unlike wrinkle ridges, high-relief ridges are symmetrical in profile and have up to 2.0 km of topographic relief; they can locally transition into lobate scarps. They were first observed in Mariner 10 images and are so-called because they were
described as ridges with ‘significant relief’ in the intercrater plains (Dzurisin, 1978). Like wrinkle ridges, high-relief ridges are found within intercrater plains and appear to have formed at approximately the same time as wrinkle ridges (Watters and Nimmo, 2010). Although these features are high-relief, they are only visible in low-sun-angle images or in Mercury Laser Altimeter data (Solomon et al., 2008; Watters et al., 2009a). These observations suggest that high-relief ridges are high-angle reverse faults.

Extensional features are rare on Mercury and were only identified in the eastern half of the Caloris basin from Mariner 10 data (Strom et al., 1975). As of this writing, MESSENGER images reveal linear to arcuate troughs, interpreted as graben, associated with two of the largest impact basins (Caloris and Rembrandt) and three intermediate-sized basins (Raditladi, Rachmaninoff, and Mozart) (Byrne et al., 2013). Caloris contains radial and concentric graben (Murchie et al., 2008; Watters et al., 2009b); Raditladi basin (255 km diameter; 27° N, 119° E) contains concentric graben; and Rembrandt basin (700 km diameter; 33° S, 88° E) contains radially oriented fractures and troughs.

Caloris basin contains Pantheon Fossae (Figure 16), a swarm of radiating troughs or graben (Murchie et al., 2008; Solomon et al., 2008; Watters et al., 2009b). On Earth, radiating graben are formed when a mantle plume or diapir impinges on the lithosphere, causing uplift and intrusion of radial dikes (e.g., Ernst et al., 1995); graben form above the rising dikes. By analogy, Pantheon Fossae and the extensional features observed in Rembrandt and Raditladi basins are interpreted to be the result of magmatic uplift sometime after basin formation (similar to floor-fractured craters on the Moon; Wichman and Schultz, 1995). Alternatively, the emplacement of volcanic plains materials outside of the basin may have locally loaded the lithosphere, causing flexural uplift – and resulting extension – in the basin interior (Melosh and McKinnon, 1988). Watters et al. (2005, 2009b) supported lateral flow of the lower crust toward the basin interior, caused by differences in elevation and crustal thickness.

Interestingly, morphologically similar trough (interpreted to be graben) swarms have been identified in Mercurian volcanic plains (Watters et al., 2012b) and are found within wrinkle ridge rings. These swarms, combined with wrinkle ridges, are interpreted to mark the locations of buried impact basins. The juxtaposition of graben encircled by compressional wrinkle ridges likely resulted from a combination of extensional stresses (cooling and thermal contraction of the plains materials) and compressional stresses (related to cooling and contraction of Mercury’s interior (Watters et al., 2012b)).

### 10.09.6.2 Mercury Volcanism

Unlike the other terrestrial planets, volcanic edifices are rare on Mercury. Head et al. (2008) characterized a possible volcanic shield volcano within Caloris basin (see also Murchie et al., 2008); Prockter et al. (2010) identified a likely volcanic vent within Rachmaninoff basin (290 km diameter; 28° N, 58° E). Hurwitz et al. (2013) identified possible lava channels on Mercury, although it is not clear if they are constructional or erosional features. The dominant expression of volcanism on Mercury, however, appears to be volcanic plains (Denevi et al., 2013; Head et al., 2008).

Head et al. (2008) examined data collected on Caloris basin from the first MESSENGER flyby and identified volcanic vents and deposits along the southern margin of the interior of the Caloris basin (Figure 17). For example, they pointed to a ‘kidney-shaped’ depression about 20 km long surrounded by relatively a relatively high-albedo deposit covering a roughly

![Figure 16](image1.png)

**Figure 16** A portion of Pantheon Fossae, one of the few extensional networks on Mercury. Image centered at 25.2° N, 158.9° E. Image acquired by the Mercury Dual Imaging System wide-angle camera; image ID 2213443. Image courtesy of NASA/JHUAPL (see [http://messenger.jhuapl.edu/gallery/sciencePhotos/pics/EW0250989390G.map.web.png](http://messenger.jhuapl.edu/gallery/sciencePhotos/pics/EW0250989390G.map.web.png)).

![Figure 17](image2.png)

**Figure 17** Likely volcanic vents (irregularly shaped depressions in the left of the image) within the rim of Caloris basin. Image centered at 22.0° N, 146.4° E. Image courtesy of NASA/JHUAPL (see [http://science1.nasa.gov/media/medialibrary/2008/07/03/03jul_mercuryupdate_resources/kidney.jpg](http://science1.nasa.gov/media/medialibrary/2008/07/03/03jul_mercuryupdate_resources/kidney.jpg)).
circular area up to ~50 km from the depression. They interpret this to be a volcanic construct (Prockter et al., 2010). Smooth plains materials in and around Caloris basin are also interpreted to be volcanic (Murchie et al., 2005); recent studies (Ernst et al., 2013; Klimczak et al., 2013) suggest that the volcanic fill within the Caloris basin is about 2.5–4.0 km thick.

There is a strong morphological similarity between smooth plains deformed by wrinkle ridges on Mercury and on Mars: both planets display intersecting wrinkle ridges systems and ridge rings (Figure 15; Head et al., 2008). The morphological similarities between lunar mare-type wrinkle ridges (found only in the mare basaltas), martian wrinkle ridges, and mercurian wrinkle ridges support the interpretation that many of the plains on Mercury are volcanic (Denevi et al., 2013).

10.09.7 Summary

The Earth is the only planetary body in the solar system that displays ‘plate tectonics,’ although Mars may have experienced Earth-like seafloor spreading briefly shortly after planetary formation. In the absence of plate tectonics, the other planetary bodies of the inner solar system display faults and fractures formed by impact- or volcanic-induced stresses or by more global stresses associated with core formation and cooling. Mare-type wrinkle ridges, first identified and characterized on the lunar basaltas, are near-surface thrust faults and are the most abundant tectonic feature in the inner solar system.

Volcanism in the inner solar system is dominated by basalt. This is the case even on Earth, where the entire ocean floor is composed of basaltic lava flows. Volcanic behaviors differ widely on the terrestrial planets, from enormous shield volcanoes on Mars to dark mantling deposits on the Moon. The longest channel in the solar system is a lava-formed channel on Venus: Baltis Vallis is longer than the Earth’s Nile River.

As we encounter these volcanic and tectonic extremes, it is essential to remember that Earth provides us with only one datum as we seek to understand the processes that shape our world. Only by examining and understanding volcanism and tectonism throughout the solar system can we be truly sure we understand how those processes operate at home.

References


Aharonson O, Zuber MT, and Rothman DH (2001) Statistics of Mars’ topography from plains on Mercury are volcanic (Denevi et al., 2013).


