Mercury's Global Evolution

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Abstract

MESSENGER's exploration of Mercury has revealed a rich and dynamic geological history and provided constraints on the processes that control the planet's internal evolution. That history includes resurfacing by impacts and volcanism prior to the end of the late heavy bombardment and a subsequent rapid waning of effusive volcanism. MESSENGER also revealed a global distribution of thrust faults that collectively accommodated a decrease in Mercury's radius far greater than thought before the mission. Measurements of elemental abundances on Mercury's surface indicate the planet is strongly chemically reduced, helping to characterize the composition and manner of crystallization of the metallic core. The discovery of a northward offset of the weak, axially aligned internal magnetic field, and of crustal magnetization in the planet's ancient crust, places new limits on the history of the core dynamo and the entire interior. Models of Mercury's thermochemical evolution subject to these observational constraints indicate that mantle convection may persist to the present but has been incapable of significantly homogenizing the mantle. These models also indicate that Mercury's dynamo generation is influenced by both a static layer at the top of the core and convective motions within the core driven by compositional buoyancy.

19. Mercury's Global Evolution

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19.1 Introduction

From formation to quiescence, the history of a planet is the consequence of an intricate set of relationships between processes that both shape the surface and operate through the entirety of the planet (Kaula, 1975). MESSENGER, which completed the first orbital investigation of Mercury in April 2015 (Chapter 1), has revealed that planet to be as rich an example of that intricacy as any of the major bodies of the inner Solar System. Mercury has long been known as a planet of enigmas, from its 3:2 spin–orbit resonance with the Sun, to its global contraction, to its unexpected magnetic field (Solomon, 2003). Now, MESSENGER has unveiled the majority of the planet that was previously unseen (Chapters 6, 9–13), characterized the large-scale chemical composition and heterogeneity of the surface (Chapters 2, 7–8), determined Mercury's shape, gravity, and rotational state (Chapters 3–4), and revealed unknown structure and ancient activity of the magnetic field (Chapter 5).

The broad set of observations of Mercury's surface and interior by MESSENGER places fundamental constraints on the processes governing the planet's evolution. Although few of these observations individually lead to unique conclusions about the history of the innermost planet, taken as a whole, and in combination with an understanding of the processes that operate on and within planets in general, they provide an important picture of how Mercury evolved. At its most basic level, a planet seen today is the consequence of how material and heat are transported on and to its surface and within the interior. Mercury's early history was marked by both intense bombardment and widespread volcanism (Chapters 6, 9, 11). Generally overprinting this record of crustal growth and reworking is a global set of tectonic features, predominantly shortening in nature and indicative of substantial contraction of Mercury, formed largely since the end of the period of heaviest bombardment of the planet (Chapter 10). MESSENGER's observations of remanent crustal magnetism during its final year in orbit revealed that Mercury possessed an internal magnetic field early in the planet's history (Chapter 5). This result indicates that within the first several hundred million years of Mercury's history, the deep interior where the magnetic field was generated was vigorously active. Each of these findings is set against the backdrop of a geochemically diverse and quite surprising surface and, by inference, interior composition (Chapters 2, 7). Indeed, MESSENGER found Mercury to be the most chemically reduced terrestrial planet on the basis of its low surface abundance of iron and relatively large surface abundance of sulfur (Nittler *et al.*, 2011). Furthermore, MESSENGER observations showed the planet to be unexpectedly volatile rich, including considerable abundances of the heat-producing elements potassium, thorium, and uranium (Peplowski *et al.*, 2011). The chemically reduced interior has major implications for the composition of Mercury's core, its structure, and how the magnetic field is generated, as does the newly constrained understanding of the abundance of heat-producing elements, which control the rate at which the planet cooled and its ability to generate magma.

In order to better understand how Mercury evolved over the past 4.5 billion years we synthesize observations by MESSENGER that elucidate the primary processes that have governed its history. We begin by outlining results from MESSENGER that clarify both how the crust of the planet formed and the history of the crust and lithosphere, including constraints from observations of surface geochemistry, the record of volcanism and tectonics, and the structure of the crust. Then we focus on observations that provide information on the state, structure, and behavior of the deeper interior. In tandem, we investigate the thermochemical evolution of the interior of Mercury subject to the constraints provided by MESSENGER's observations. Finally, we discuss the implications of these results for the history of the planet and outline prospects for future progress on understanding how the whole of Mercury has evolved.

19.2 Earliest history of the crust

19.2.1 Geological constraints

The geologic record of Mercury's earliest crust – the outermost, petrologically distinct layer of the silicate portion of the planet derived from melting of the mantle (e.g., Brown and Elkins-Tanton, 2009; Namur et al., 2016; Namur and Charlier, 2017; Chapter 3) – is largely obscured by resurfacing by both impacts and volcanism (e.g., Trask and Guest, 1975; Spudis and Guest, 1988; Strom and Neukum, 1988; Denevi *et al.*, 2009). Indeed, the most heavily cratered terrain has been estimated to have an age of 4.0–4.1 Gyr (Marchi *et al.*, 2013). However, despite the fact that there are no areas of the crust that can be quantifiably ascribed to the first ~500 Myr of Mercury's history, important clues to the nature and origin of the crust are found in several areas that appear to have undergone only minimal resurfacing as well as in material exposed from depth by large impact events (Chapter 6).

Spectral units termed low-reflectance material (LRM) (Robinson *et al.*, 2008; Denevi *et al.*, 2009; Murchie *et al.*, 2015; Klima *et al.*, 2016) appear to be one key to our understanding of Mercury's crust. With a reflectance of just 4–5% at 550 nm (Chapter 8), LRM is ~30% darker than Mercury's average surface and is found concentrated in the ejecta of large impact craters (Denevi *et al.*, 2009; Ernst *et al.*, 2010; Klima et al., 2016). The reflectance and spectral properties of the LRM are consistent with the deposits having a graphite component (Murchie *et al.*, 2015). Furthermore, increases in thermal neutron count rates associated with LRM deposits suggest a carbon abundance that is 1–3 wt% higher than that of surrounding terrain (Peplowski *et al.*, 2015a, 2016). These observations are consistent with the hypothesis that Mercury developed a carbon-rich floatation crust due to buoyancy of graphite in an early magma ocean (Vander Kaaden and McCubbin, 2015).

Any early crust, particularly one as thin as a graphite-rich crust might have been, was surely disrupted heavily by impacts, modified by magmatic intrusions, and buried by volcanic deposits. Therefore, the modern distribution of this primordial material on the surface is limited, as it has been substantially mixed and diluted with other materials. By this reasoning, LRM is the material with the greatest concentration of carbon in a C-rich crust (Peplowski *et al.*, 2015a). The depth of origin of LRM, calculated from the excavation depth of impact craters, is often several to tens of kilometers (Denevi *et al.*, 2009; Ernst *et al.*, 2010; Ernst *et al.*, 2015; Peplowski *et al.*, 2015a). These depth estimates provide lower bounds to the depth of burial by impact and volcanic deposits subsequent to the formation of the original floatation crust. In some of the most heavily cratered terrains, the overall surface is relatively low in reflectance and all impact craters in the region expose LRM, suggesting that these regions may have experienced less resurfacing than average (Chapter 6). However, in other large regions, no LRM is found in any crater smaller than ~150 km in diameter, suggesting burial by at least 8 km of volcanic material (Chapter 6). Rivera-Valentin and Barr (2014) explored impact redistribution models for an impactor population consistent with Mercury's cratering record and found that the LRM is consistent with a darkening agent approximately 30 km deep, which would be within the lowermost crust or upper mantle (James *et al*., 2015; Padovan *et al*., 2015). Concentration of a darkening agent, such as graphite, from a crustal layer deep within the crust may also imply that volcanism was substantial and occurred with a flux much greater than impact redistribution of upper crustal material in the period before the onset of the late heavy bombardment (LHB).

Otherwise, the darkening agent would have been efficiently mixed throughout the crust and unlikely to display variations associated with exhumation from depth.

19.2.2 Geochemical state of the crust and mantle

The composition and chemical diversity of the surface of Mercury provide important insights into the nature, origin, and evolution of the crust and mantle. Given that Mercury is strongly differentiated with an uncommonly low silicate-to-metal ratio (Chapters 2, 4), understanding the mechanisms that may be responsible for Mercury's crustal formation has been a long-standing question. At their most basic, models for the formation of the crust include partial melting of an undifferentiated, chondritic-like mantle; formation as the uppermost layer of a solidifying magma ocean; or products of remelting of a magma ocean.

Geochemical observations and the relative ages of the surface units of Mercury argue against an undifferentiated mantle as the source region for melts erupted onto the surface. Melting of enstatite chondrites has been investigated experimentally and modeled from phase equilibria to understand both the origin of the highly-reduced aubrite parent body (McCoy *et al.*, 1999) and Mercury (Burbine *et al.*, 2002; Malavergne *et al.*, 2010). An undifferentiated chondritic mantle would produce sodium-rich melts at low degrees of partial melting, consistent with the composition of the northern smooth plains (NSP) (Vander Kaaden and McCubbin, 2016). However, the high Mg/Si and low Al/Si ratios observed for Mercury's average surface composition require relatively high degrees of partial melting (Burbine *et al.*, 2002; Nittler *et al.*, 2011). Further, the formation of the high-sodium flood basalts of the NSP relatively late in the history of Mercury would require a fertile mantle source that had not experienced earlier partial melting. Finally, the highly differentiated nature of Mercury, including the presence of a large core, argues against preservation of a wholly undifferentiated mantle.

A widely accepted model of the mantle and crust suggests that Mercury once had a magma ocean responsible for an initial stage of silicate differentiation. Prior to MESSENGER's orbit insertion and the early geochemical measurements of the surface of Mercury, the nature of the crust and the bulk composition of the surface and planet were poorly constrained, although the surface was known to be FeO-poor and the bulk composition of the planet rich in iron metal (Taylor and Scott, 2003, and references therein). This uncertainty led to a range of magma ocean models producing either a plagioclase floatation crust or a low-FeO magmatic crust, depending on the bulk composition of the magma ocean (Brown and Elkins-Tanton, 2009; Riner *et al.*, 2009). Some of these petrologic models produce gravitationally unstable mantles that would experience overturn, similar to that posited for the lunar mantle.

With the realization that the crust of Mercury is neither a plagioclase-rich floatation crust nor chemically homogeneous, models emerged that considered a magma ocean with subsequent remelting (Charlier *et al.*, 2013; Vander Kaaden and McCubbin, 2015, 2016). Charlier *et al.* (2013) suggested that compositional heterogeneity observed during early MESSENGER orbital observations could have been the result of melting of different layers within the mantle during convection and adiabatic pressure-release melting, even in the absence of mantle overturn. Vander Kaaden and McCubbin (2015) strengthened the argument against a significant primary floatation crust experimentally by demonstrating that graphite is the only phase that is buoyant in a Mercury magma ocean. The equivalent thickness of such a graphite layer is directly dependent on the concentration of carbon in the silicate portion of the planet. Should Mercury have a bulk silicate carbon content similar to those of Earth, Mars, or the Moon, that layer might be up to

~100 m thick. However, if Mercury's carbon content is more similar to that of chondritic materials, a graphite crust could range in thickness from as little as 100 m to more than 10 km, with the largest values for bulk silicate compositions similar to carbonaceous chondrites (Vander Kaaden and McCubbin, 2015). These authors further noted that, unlike many other planetary bodies, partial melts derived from mantle melting on Mercury are buoyant throughout the mantle and would rise to the surface without stalling at some neutral buoyancy depth. Thus, the crust of Mercury is likely comprised of an impact-gardened mixture of primary crust formed during a magma ocean stage and subsequent volcanic deposits. Vander Kaaden and McCubbin (2016) further refined this idea by noting that a crystallizing magma ocean without buoyant silicate phases would concentrate incompatible elements, including volatiles, near the surface of the planet. Thus, remelting of shallow cumulates can produce volatile-rich compositions, like the NSP, even at high degrees of partial melting.

19.3 History of the crust and lithosphere

Geological observations provide compelling evidence that Mercury's crust is largely volcanic in origin and has experienced widespread tectonic deformation. The accumulated, observable history of Mercury's crust and lithosphere contains fundamental clues to the processes that shaped the surface of the planet, and importantly, the time progression of these processes. Whereas the earliest history of the planet may have included a magma ocean and the generation of a thin and rather exotic floatation crust, it is the subsequent history that is more discernable. MESSENGER's collected geophysical, geological, and geochemical observations of Mercury provide important insights into both the planet's integrated history as well as many discrete events, of variable duration, that reflect its evolutionary path.

19.3.1 Crustal thickness

In addition to the geochemical and geological markers of crustal formation, Mercury's gravity field and topography provide important clues to the nature and formation of the crust (Perry *et al.*, 2015; Tosi *et al.*, 2015; Chapter 3). Mercury's crust is the product of the combined processes of crystallization of any magma ocean and upward transport of mantle partial melts integrated over the course of the planet's history. Therefore, knowledge of the thickness of the crust is a crucial indicator of the efficiency and pattern of igneous differentiation of the planet, which in turn depend strongly on Mercury's internal activity.

Orbital observations of Mercury's gravity field by MESSENGER provided the first detailed measurements of its mass distribution. MESSENGER's eccentric orbit (Chapter 1), with the periapsis located at a high northern latitude, resulted in gravity field measurements that have the highest spatial resolution in the north and that resolve only much longer wavelengths in the southern hemisphere (Smith *et al.*, 2012; Mazarico *et al.*, 2014; Verma and Margot, 2016). Focusing on the higher-resolution information in the northern hemisphere, several estimates of the thickness of the crust have been calculated (Smith *et al.*, 2012; James *et al.*, 2015; Padovan *et al.*, 2015). The most recent of these models place the average crustal thickness of the northern hemisphere at 35 ± 18 km on the basis of geoid-to-topography ratios (GTR) (Padovan *et al.*, 2015) and place a minimum on the average thickness of 38 km with a model that accounts for both crustal and mantle sources of compensation (James *et al.*, 2015). Density differences between the crust and mantle are a major source of uncertainty in crustal thickness models. Padovan *et al.* (2015) considered a range of crustal densities from 2700 to 3100 kg m^{-3}, with the upper bound consistent with grain densities they inferred from MESSENGER elemental compositions, the lower bound the result of including 12% porosity throughout the crust, as has been inferred for the Moon (Wieczorek *et al.*, 2013), and a mantle density of 3300 kg m^{-3}. This range overlaps independent estimates of the grain densities calculated from experimental determinations of the modal mineralogy consistent with the range of surface compositions across Mercury (Namur and Charlier, 2017). Similarly, the inversion approach of James *et al*. (2015) was for a nominal crustal density of 3200 kg m^{-3} and a mantle density of 3400 kg m^{-3}. Generally speaking, the small difference in grain density between the crust and mantle, approximately 200 kg m^{-3}, is a reflection of the inferred low iron content of Mercury's silicate layers. This density difference is also important for crustal flow models, as the driving stress for any topographic relaxation via lower crustal flow scales directly with the density contrast (e.g., Nimmo and Stevenson, 2001), so a small density contrast implies less lower crustal flow. Potentially of greater importance is that the inferred crustal thickness values when compared with the thickness of the mantle imply that Mercury has experienced the most efficient extraction of crust among the terrestrial bodies. Indeed, Mercury's crust represents approximately 10% of all silicate material on the planet (James *et al.*, 2015; Padovan *et al.*, 2015). Such efficient extraction is likely the result of relatively high degrees of partial melting, consistent with geochemical observations of the surface and inferences for the interior (Chapters 2, 7).

Compared with Mercury's global shape as derived from laser altimetry and radio occultation measurements, the geoid has a spectral power of only ~1% that of the shape at spherical harmonic degree and order two, which indicates that topographic variations on Mercury at the longest wavelengths are largely isostatically compensated (Perry *et al.*, 2015). Should the variations at degree and order two be compensated by variations in the thickness of the crust, this difference would imply a ~24 km pole-to-equator change in crustal thickness. However, other mechanisms such as variations in density due to temperature or composition may contribute to the compensation, potentially reducing any long-wavelength crustal thickness variation (Perry *et* *al.*, 2015; Tosi *et al.*, 2015; Chapter 3). Regardless, a substantial latitudinal variation in the crustal thickness of Mercury would be an important, if as yet poorly understood, constraint on crustal production (Chapter 3).

19.3.2 Surface history

One of the more direct measures of the evolution of a planet's crust is the geological history of its surface. To first order, Mercury's surface can be classified into units of either smooth plains or intercrater plains (Chapter 6). The former type of unit is texturally smooth and relatively sparsely cratered, displays sharp boundaries with adjacent regions, and is level to gently sloped over baselines of ~100–200 km (Trask and Guest, 1975; Denevi *et al.*, 2013a; Chapter 6). These smooth plains units occupy about 27% of the planet's surface (Figure 19.1) and are predominantly located in the northern hemisphere in the NSP and within and adjacent to the Caloris basin. The remainder of the surface is largely dominated by intercrater plains, which are characterized by gently rolling terrain with gradational boundaries and a greater density of secondary craters 5–10 km in diameter than smooth plains (Trask and Guest, 1975; Denevi *et al.*, 2013a). The intercrater plains are situated between individual and clusters of large (>30 km) craters, which generally superpose the plains and are the source of the secondary craters. As the density of superposed impact craters appears to be the main distinction between the varieties of plains (Byrne *et al*., 2016), their main difference likely reflects a range in age rather than specific lithological or rheological differences (Murray *et al.*, 1975; Strom, 1977; Spudis and Guest, 1988; Denevi *et al.*, 2009; Whitten *et al.*, 2014). Little evidence remains of an older, more heavily cratered surface apart from several regions that have undergone only partial resurfacing or portions of basin massifs that predate the intercrater plains (Chapter 6).

Observations of Mercury have established that the planet has been heavily shaped by volcanic activity. For example, the majority of smooth plains units are interpreted as effusive volcanic deposits, on the basis of their distinct unit boundaries, embayment relations with surrounding topography, the presence of buried "ghost craters" within these units, spectral differences with neighboring terrain, and deposits located far from any large basins (Murray *et al.*, 1974, 1975; Strom *et al.*, 1975; Spudis and Guest, 1988; Robinson and Lucey, 1997; Head *et al.*, 2008, 2011; Murchie *et al.*, 2008; Robinson *et al.*, 2008; Denevi *et al.*, 2009, 2013a; Chapter 11). A number of other volcanic landforms formed by effusive activity have also been reported across the planet, including a small shield volcano, lobate flow margins, and lava-sculpted valles (Head *et al.*, 2008, 2011; Byrne *et al.*, 2013; Hurwitz *et al.*, 2013). Landforms attributed to explosive volcanism (e.g., Kerber *et al.*, 2009; Thomas *et al.*, 2014), often in close spatial proximity to smooth plains, have also been identified.

The major smooth plains deposits on Mercury have crater densities that vary by up to a factor of 5 for craters larger than 10 km. However, because of the inferred rapid decline in cratering during their formation, their derived model ages are the same, within statistical error, for any of the published model production function (MPF) chronologies for Mercury (Strom and Neukum, 1988; Neukum *et al.*, 2001; Marchi *et al.*, 2009; Le Feuvre and Wieczorek, 2011), though differences among the model chronologies are greater for lower-density (younger) deposits (Chapter 9). Crater size–frequency analyses have shown that the NSP, the single largest smooth plains deposit on the planet (Chapter 6), as well as the plains interior to the Caloris and the Rembrandt impact basins, were emplaced around 3.7–3.8 Ga (Fassett *et al.*, 2009; Head *et al.*, 2011; Strom *et al.*, 2011; Denevi *et al.*, 2013a; Ferrari *et al.*, 2014; Ostrach *et al.*, 2015; Chapter 9). The areal densities of impact craters for two additional large smooth plains deposits on

Mercury, those near the Faulkner crater and the Rachmaninoff basin, are comparable to the densities for the NSP and Caloris interior plains (Fassett *et al.*, 2009; Denevi *et al.*, 2013a; Whitten *et al.*, 2014; Ostrach *et al.*, 2015), implying that these other units are similar in age. Crater density measurements for several additional, smaller smooth plains deposits yield ages of ~3.8–3.5 Ga for these sites (Byrne *et al.*, 2016) with the crater model production function of Le Feuvre and Wieczorek (2011). Only one definitively volcanic smooth plains deposit has been identified on the planet with a substantially younger age than those above. Situated within the inner peak ring of the Rachmaninoff impact basin, this deposit is considerably smaller than other plains units for which ages have been determined (Prockter *et al.*, 2010; Marchi *et al.*, 2011). The distribution of the model ages of smooth plains units (in particular the units shaded dark purple in Figure 19.1), which are stratigraphically the youngest effusive volcanic features on Mercury, suggest therefore that flood volcanism was largely completed by ~3.5 Ga (Byrne *et al.*, 2016).

Similar to the smooth plains, intercrater plains units have a range of crater areal densities, the lowest values of which overlap the highest corresponding values for smooth plains units (Whitten *et al.*, 2014; Byrne *et al.*, 2016). The model ages of the intercrater plains are ~3.9–4.1 Ga (e.g., Whitten *et al.*, 2014; Chapter 9). Notably, nowhere on Mercury is as heavily cratered as the lunar highlands (Strom, 1977; Strom *et al.*, 2008; Fassett *et al.*, 2011; Marchi *et al.*, 2011), and the most heavily cratered regions on Mercury have been dated at just 4.0–4.1 Ga (Marchi *et al.*, 2013) with the chronology of Marchi *et al*. (2009). These model age results suggest that little remains of the geologic record of the earliest ~500 Myr of Mercury's surface history (Chapters 6, 9).

The origin of Mercury's intercrater plains is less certain than that of the smooth plains, but they may also be dominantly products of volcanism. The main line of evidence lies in their age: model ages of 4.0–4.1 Ga require major resurfacing of the earliest crust, with volcanism being a likely major cause (Head *et al.,* 2011; Denevi *et al.*, 2013a; Whitten *et al.*, 2014; Chapters 6, 11). For example, Whitten *et al*. (2014) showed that cratering of smooth plains, particularly by secondaries from nearby primary craters, renders those smooth deposits texturally similar to intercrater plains. Large regions within the intercrater plains have also been interpreted as volcanic in origin on the basis of a substantial deficit of the most degraded class of craters, as well as stratigraphic and color relationships that are analogous to volcanic smooth plains deposits (Denevi *et al*., 2013b; Chapter 6). Although discrete volcanic landforms may not have survived the history of impact bombardment of Mercury prior to the emplacement of the smooth plains, thermochemical evolution models of the planet imply that voluminous and widespread effusive volcanic activity operated for at least the planet's first half-billion years (Michel *et al.*, 2013; Tosi *et al.*, 2013). If so, then the intercrater plains we observe today are likely just older smooth plains deposits (e.g., Strom, 1977; Spudis and Guest, 1988; Denevi *et al.*, 2009; Whitten *et al.*, 2014). This inference is consistent with the observed compositional heterogeneity on Mercury, where differences in composition do not always follow morphologic boundaries, and where smooth and intercrater plains can share similar compositions (Weider *et al.*, 2015).

The cessation of large-scale effusive volcanism on Mercury, as seen in the smooth plains and the older intercrater plains, effectively heralded the end of the crust-building phase of Mercury's evolution, but volcanic activity in some form continued thereafter. For example, the identification of irregular pits across Mercury, often characterized by a lack of a raised rim, scalloped edges, and diffuse-edged deposits with a distinct reddish color, provides evidence for explosive volcanism having occurred on the planet (Head *et al.*, 2008; Murchie *et al.*, 2008; Kerber *et al.*, 2009; Chapter 11). Some of these pyroclastic deposits may be as young as ~1 Ga (Thomas *et al.*, 2014). Many of Mercury's explosive volcanic landforms and deposits are spatially associated with areas of pre-existing crustal weaknesses, including the surface breaks of thrust faults underlying lobate scarps and within the heavily fractured central peaks and peak rings of craters (Figure 1) (Kerber *et al.*, 2011; Thomas *et al.*, 2014; Chapter 10). Additionally, the areal extents of pyroclastic deposits are far less than those of effusive volcanic deposits. Although widely distributed, the role of explosive volcanism in the building and resurfacing of Mercury's crust was negligible compared with the contribution from effusive volcanism.

The history of Mercury's surface is recorded as much in its tectonic landforms as in its volcanic ones. Indeed, the surface of Mercury is replete with tectonic features, including landforms termed "wrinkle ridges" and "lobate scarps" (see the bottom panel of Figure 1), interpreted to have accommodated crustal shortening in response to global contraction (Strom *et al.*, 1975). The number and structural relief of this ensemble of landforms correspond to a decrease in planetary radius of at least 5 to 7 km (Byrne *et al.*, 2014; Chapter 10). These figures are in stark contrast with earlier estimates from more limited Mariner 10 data and early flyby data from MESSENGER that suggested that perhaps no more than 2 km of contraction was likely (Strom *et al.*, 1975; Watters *et al.*, 1998, 2009). Importantly, crater and thrust fault superposition relations indicate that global contraction was underway by around the time that widespread effusive volcanism came to an end (Banks *et al.*, 2015; Byrne *et al.*, 2016). Observations of craters formed during the Calorian system (Spudis and Guest, 1988; Chapter 9) that superpose scarps show that shortening of Mercury's surface on at least a regional scale had begun at some time before ~3.6 Ga (Banks *et al.*, 2015). Further, the discovery with

MESSENGER low-altitude image data of a population of lobate scarps at least an order of magnitude smaller than previously recognized (Watters *et al.*, 2015b), and the stratigraphic relationships between such scarps and impact craters with a range of degradation states, is suggestive that tectonic accommodation of global contraction persisted over most of Mercury's history (Banks *et al.*, 2015).

Observations made with MESSENGER data have helped characterize the resurfacing mechanisms and history of the innermost planet. Voluminous magma genesis within Mercury's interior likely resulted in globally extensive effusive volcanism that persisted for at least several hundred million years. This volcanic activity, together with an increase in the impact flux at the start of the LHB, has obscured the geological record of the first ~500 Myr of Mercury's surface history. With a reduction in magma genesis as a result of secular cooling and with the horizontal compressive state in Mercury's lithosphere resulting from global contraction, widespread effusive volcanism began to wane, with eruptive volumes decreasing with time, before ultimately ending by about 3.5 Ga. Explosive volcanism endured for far longer, but the vast majority of Mercury's crust was in place prior to 4 Ga, and smooth plains formation constituted the tapering end of the planet's crust-building phase.

19.3.3 Chemical and petrological constraints on crustal formation

Observations by MESSENGER's suite of geochemical sensors have provided important insight into the composition of the planet, the make-up of the crust, and how it formed (Chapters 2, 7). In particular, the X-Ray Spectrometer (XRS), Gamma-Ray Spectrometer (GRS), and Neutron Spectrometer (NS) provided spatially resolved surface abundances of U, K, and Th, as well as Si-normalized elemental abundances for Na, Mg, Al, S, Cl, Ca, Ti, Cr, Mn, Fe, and O. On a global scale, XRS measurements (Nittler *et al.*, 2011) indicate that the surface of Mercury exhibits a high Mg/Si ratio (0.33–0.67), which is intermediate between those of terrestrial oceanic and lunar mare basalts and highly magnesian komatiites. Mercury's surface also exhibits lower Al/Si and Ca/Si ratios than typical terrestrial or lunar basalts. Most surprising, high S/Si ratios (0.05–0.15) suggest abundances of the moderately volatile element S up to ~4 wt %. Observations from the GRS further argue against a volatile-depleted composition for Mercury. For example, Mercury's K/Th ratio is comparable with that of other terrestrial planets and is much higher than observed in the volatile-depleted lunar crust (Peplowski *et al.*, 2011). Moreover, large ratios of Na/Si (0.12) and Cl/Si (0.0057) are also observed (Evans *et al.*, 2012, 2015). Together, these observations suggest a magnesium-rich, iron-poor crust formed under chemically reducing conditions, yet not depleted in volatiles as had been predicted for an iron-rich planet so close to the Sun (e.g., Taylor and Scott, 2003).

The surface of Mercury exhibits considerable chemical and, therefore by extension, mineralogical diversity. This diversity is best documented in the northern hemisphere, where high-spatial-resolution measurements allow us to distinguish discrete geochemical terranes (Chapters 2, 7). These include the northern geochemical terrane, the Caloris interior plains terrane, the high-magnesium terrane, and the "low fast" terrane (so named because it has a low count rate for fast neutrons). Among these terranes, the northern geochemical terrane and the low fast terrane are present largely, though not exclusively, within the northern smooth plains. The Caloris interior plains terrane corresponds spatially to the boundaries of the smooth plains within the Caloris impact basin. In contrast, the high-magnesium terrane is geochemically coherent in a number of features but exhibits no clear correlation with spectral or morphometric features across the entirety of the region. However, while the crustal thickness within the majority of the region is similar to the average of the northern hemisphere, the northern and eastern boundaries are approximately coincident with areas that transition from average to thicker-than-average crust (Chapter 3). In contrast to the well-resolved XRS measurements in the northern hemisphere, the large XRS footprints in the southern hemisphere yield only a single hemispheric average composition.

Chemical compositions derived from the four distinct, northern-hemisphere geochemical terranes range in composition from basaltic andesite to trachyte on the basis of their total alkali content (Na and K) compared with silica, but ultimately all share a boninite classification due to their high MgO (> 8 wt %) and low TiO_2 (<0.5 wt %) concentrations. A common feature of all these geochemical terranes is that each has high volatile element concentrations, with Na ranging from 2.6 to 5.7 wt % and S from 1.8 to 2.9 wt %. Considerable geochemical differences do exist between the terranes, particularly with respect to Na, Mg, Al, and Fe, all of which differ by factors of 1.8 or greater among the terranes. The low fast terrane is most similar to the average surface composition for the planet. However, it is geochemically distinct from the northern geochemical terrane, with the two terranes combined occupying much of the NSP. Mineralogically, these terranes share the common feature of being unusually rich in normative plagioclase (37–58 wt %) (see Chapter 7). If classified as plutonic igneous rocks, these terranes would include norite, anorthositic norite, and anorthositic gabbro, reflecting differences in plagioclase abundance and the ratio of high-calcium to low-calcium pyroxene. The high-magnesium terrane is distinctive in its unusually high concentration of normative olivine (31 wt %).

19.3.4 Evolution of the lithosphere

The mechanical behavior of the lithosphere – the outer portion of the planet that behaves as a mechanically strong layer and includes portions of the crust and possibly mantle – provides key insights into the history of planetary stresses and temperatures. This behavior is recorded in both the tectonic landforms the lithosphere hosts (e.g., Chapter 10) and in the flexural response to loads inferred from gravity and topography data (e.g., Chapter 3). For example, elastic dislocation modeling of topographic profiles derived from Mariner 10 stereophotogrammetric data of select lobate scarp features indicates that the underlying faults penetrate, and thus the lithosphere deformed in a brittle fashion, to depths of 25–30 km, at the time of faulting (e.g., Watters and Nimmo, 2010). A comprehensive assessment of the spatial variation in lithospheric thickness from this or similar techniques, however, has yet to be completed. Interestingly, this estimate of the depth to the brittle–ductile transition from lobate scarp fault depths is consistent with models of lithospheric strength for time periods prior to ~3.5 Ga (Williams *et al.*, 2011).

At regional and more local scales, stresses in the lithosphere can be modified by loads produced by volcanism and impact basin formation and evolution (e.g., Kennedy *et al.*, 2008; Freed *et al.*, 2009; Blair *et al.*, 2013). Nonetheless, the history of the state of stress in Mercury's lithosphere has been dominated by some combination of two independent, globally acting processes: despinning from a likely early, rapid rate of rotation, and global changes in planetary radius arising from internal temperature changes (and cooling in particular) (Chapter 10). During planetary spin-down, an equatorial bulge supported by the planet's lithosphere would have relaxed (Melosh, 1977), forming a global set of near-surface joints with no preferred orientation at the poles but with an increasingly prominent east–west fabric toward the equator, under the assumption of a globally uniform lithospheric thickness (Klimczak, 2015). Similarly, also under the assumption of globally uniform lithospheric properties, a reduction in planetary volume from cooling of the core and mantle and from mineralogical phase changes (e.g., core crystallization) would yield a stress state in which horizontal compressive stresses exceed vertical stresses and under which a global set of thrust faults with no preferred orientation would develop (Melosh and McKinnon, 1988). However, spatial variations in lithospheric thickness, such as those imparted by long-lived latitudinal and longitudinal differences in surface temperature (Williams *et al.*, 2011; Tosi *et al.*, 2015), may further have influenced how tectonic deformation on Mercury was exhibited (Beuthe, 2010). Furthermore, the combination of stresses from both despinning and global contraction may have had a substantial influence on the stress state and the style of brittle tectonic deformation at the surface of the planet (Melosh and Dzurisin, 1978; Dombard and Hauck, 2008; Beuthe, 2010). Reorientation as a result of true polar wander, for example, driven by the formation of a large load such as the Caloris basin (Matsuyama and Nimmo, 2009), could also have altered the prevailing stress state.

Global mapping of Mercury's tectonic landforms from MESSENGER image data (Byrne *et al.*, 2014; Watters *et al.*, 2015a) has not revealed, to first order, any evidence of planet-wide, organized patterns of tectonic landforms predicted by earlier studies of despinning (e.g., Melosh, 1977; Melosh and McKinnon, 1988; Matsuyama and Nimmo, 2009; Beuthe, 2010). Given Mercury's near-zero obliquity, solar illumination at the equator is always due east or due west, which facilitates the identification of tectonic landforms that strike ~north–south (Byrne *et al.*, 2014) more easily than those trending east–west at low to mid latitudes. Landforms between 60°S and 60°N, in particular, show a predominantly north–south orientation. Landforms north of 60°N show some clustering at southwest–northeast trends but are not as strongly oriented as those at mid- to low latitudes; landforms south of 60°S show no preferred orientations (Byrne *et al.*, 2014). Nonetheless, when the effect of solar azimuthal illumination is considered, a general

~north–south trend for mid- to low-latitude landforms remains (Watters *et al.*, 2015a). Under the assumption that currently published tectonic maps are generally complete and that no bias in lighting geometry has obscured substantial ~east–west-tending landforms yet to be identified (Chapter 10), the history of stress within Mercury's lithosphere must be reconciled with these observations.

The lack of opening-mode fractures on Mercury, aside from those identified within volcanically flooded impact features (Freed *et al.*, 2012; Klimczak *et al.*, 2012; Chapter 10), indicates that no direct evidence of tidal despinning alone remains. Given that tidal despinning likely occurred geologically rapidly – although the timing of this process remains to be characterized fully – it is perhaps no surprise that such evidence is missing from the geological record, especially given that the oldest terrain on Mercury is ~4.1 Ga (section 19.3.2). On the other hand, shortening structures on the innermost planet do not form a globally heterogeneous pattern, the expected result of global contraction alone.

It may be, then, that the most straightforward interpretation of the global pattern of orientations of Mercury's tectonic landforms represents some combination of despinning and global contraction (Klimczak, 2015; Chapter 10). Thrust faults developing in such a stress state would have developed with preferred north–south orientations near the equator (Klimczak, 2015), though the expected pattern near the poles could be oriented either somewhat randomly (Klimczak, 2015), depending on rock strength, or might have a more organized pattern if latitudinal variations in lithospheric thickness were substantial (Beuthe, 2010). Regardless, this general pattern in the mid-latitudes with a differing orientation at high-latitudes is similar to that observed from global mapping (Byrne *et al.*, 2014). As such, it indicates that either tidal despinning temporally overlapped with global contraction, or that despinning imparted some fabric to Mercury's lithosphere that survived until the onset of, and influenced the tectonic deformation from, global contraction. The lack of a clear signature of reorientation stresses as would be reflected in an orientation of the lobate scarp structures (Matsuyama and Nimmo, 2009) suggests that true polar wander was not a major component of the processes that drove tectonic deformation. Thus, the relative timing of global contraction and despinning, as well as the effects of spatial variations in lithospheric thickness (e.g., Beuthe, 2010) when considered with possible values for the degree of lithospheric fracturing (e.g., Klimczak, 2015), are important questions that remain outstanding.

19.4 Knowledge of the interior

19.4.1 Constraints on core composition

Observations of low ferrous iron concentrations and larger-than-expected sulfur abundances on Mercury's surface indicate that the planet's surface and, by extension, interior are strongly chemically reduced (Nittler *et al.*, 2011; Weider *et al.*, 2012; Chapter 2). Inferred oxygen fugacities range between 3 and 7.3 $\log_{10}$ units below the iron–wüstite (IW) buffer, with a consistent overlap between published estimates of IW-4.5 to IW-6.3 (Malavergne *et al.*, 2010; McCubbin *et al.*, 2012; Zolotov *et al.*, 2013). Under these highly reducing conditions, elements that are normally lithophile and incorporated into silicates and oxides can instead have chalcophile or siderophile behavior, combining to form sulfides or metallic phases. This behavior is observed in aubrite meteorites, which formed at similarly reducing conditions and contain a host of exotic sulfides, metals, carbides, phosphides, and nitrides (Keil, 1989). For the surface of Mercury, correlated S/Si and Ca/Si abundances have been invoked to postulate the presence of oldhamite (CaS) (Nittler *et al.*, 2011), although more recent studies (Stockstill-Cahill *et al.*, 2012; Vander Kaaden and McCubbin, 2016; Vander Kaaden et al., 2016) favor complex sulfides of Fe, Mg, Ti, Cr, and Mn, as well as Ca.

Although Fe and Ni are expected to be primary components of Mercury's core, the highly reducing conditions inferred from surface materials, if indicative of conditions in Mercury's interior, will also have led to the incorporation of light elements into the core, most notably Si and S. As conditions become more reducing, Si becomes more soluble in the metal phase (e.g., Berthet *et al.*, 2009; Malavergne *et al.*, 2010; Chabot *et al.*, 2014). A similar trend is also observed as temperatures increase (McCoy *et al.*, 1999). During planetary differentiation, Si can be incorporated into the metallic phase and thus into the core of a planetary body. In practice, the segregated core material under reducing conditions contains S as well, so that the Fe–S–Si system rather than the binary Fe–Si system governs phase relations. Within the Fe–S–Si system (Raghavan, 1988), liquid immiscibility can occur, producing separate S-rich and Si-rich metallic liquids. Chabot *et al*. (2014) noted this behavior in experiments in which Fe–S–Si liquids occurred in equilibrium with silicate melts for which the sulfur concentration was comparable with that observed on the surface of Mercury (1–4 wt % S). These authors noted that, whereas the co-existing Fe–S–Si melts (and, by extension, the core of Mercury) can readily contain both sulfur and silicon, changing oxygen fugacity may result in either high-Si, low-S or high-S, low-Si melts, either of which could satisfy the constraint imposed by the presence of sulfur-rich silicate melts on the surface of Mercury.

Additional, less well-constrained light elements that might be incorporated into the core are C and P. The inference that graphite could both crystallize from and occur as a floatation product in an early Mercury magma ocean (Vander Kaaden and McCubbin, 2015) suggests that Mercury's core might be saturated in carbon if the mantle and core were in equilibrium, as carbon tends to be siderophile at reducing conditions. Although likely a minor constituent in Mercury's core, it may have a substantial effect on its melting behavior (Deng *et al.*, 2013a; Martin *et al.*, 2014; Martin *et al.*, 2015). Phosphorus also behaves as a siderophile under reducing conditions, forming Fe,Ni-phosphides. No measurement of phosphorus on the surface of Mercury yet exists, although its incorporation into the core in minor concentrations would be expected, potentially resulting in a complex behavior of the core governed by the Fe–S–Si–C–P system.

19.4.2 Internal structure

Planetary evolution is intimately intertwined with the distribution of materials within the interior of the body. The processes of metal–silicate differentiation, core crystallization, mantle convection, and magmatism tend to result in a layered compositional and density structure within the interior of a planet. Such a layered structure is typically comprised of one or more Fe-rich layers in the planet's core, as well as one or more silicate mantle layers, all topped by a silicate crust. The thickness, material properties, and heat-producing element content of each of these layers controls how the planet generates and loses heat, generates magma, and produces a magnetic field – although neither of the latter two is guaranteed.

Mercury's large bulk density of ~5430 km m^{-3} has long been understood to imply that the planet has an unusually large metal-to-silicate ratio (Siegfried and Solomon, 1974; Solomon, 1976). Consequently, Mercury has a relatively large metallic core with a comparatively thin layer of overlying silicate material. Mariner 10's discovery of Mercury's magnetic field (Ness *et al.*, 1975) suggested the possibility that the core could be partially molten. The presence of a liquid layer within the core was subsequently confirmed by Earth-based radar observations of the libration and orientation of Mercury (Margot *et al.*, 2007; Chapter 4). Pre-MESSENGER studies also used the long-wavelength gravity field to estimate the thickness of the planet's crust

(Anderson *et al.*, 1996) at ~100–300 km. Such a large crustal thickness was exceptionally surprising because it represented one-sixth to one-half of the estimated silicate content of the planet, far in excess of that for any other known planetary body. Constrained only by the radius and bulk density of the planet, the relative size of Mercury's core, and whether it contained a solid inner core, remained similarly uncertain. However, in models of Mercury's internal evolution it was commonly assumed that the silicate portion of the planet was ~600 km thick with the remainder of the interior comprised of an Fe-rich core (e.g., Schubert *et al.*, 1988; Hauck *et al.*, 2004; Redmond and King, 2007; Grott *et al.*, 2011).

Measurements of Mercury's gravity field by the MESSENGER spacecraft have led to greatly improved estimates of the planet's internal structure (Smith *et al.*, 2012; Hauck *et al.*, 2013; Rivoldini and Van Hoolst, 2013; Chapter 4). That Mercury occupies a Cassini state, wherein the rotation axis is approximately perpendicular to the plane of its orbit about the Sun and the spin and precession rates of the planet are equal, presents an opportunity to estimate the planet's structure. Indeed, a procedure was developed to determine the normalized polar moment of inertia and the fraction of that moment contributed by the outermost solid portion of the planet (e.g., Peale, 1988; Peale *et al.*, 2002) as a result of Mercury's special rotation state. The background and details of this experiment and its interpretation are discussed at length in Chapter 4. The fundamental result is that through measurement of just four quantities – the polar and equatorial oblateness of the gravity field expressed as the second-degree spherical harmonic coefficients C_{20} and C_{22}, the amplitude of the physical libration, and the obliquity of the planet – it is possible to resolve two measures of the radial density distribution of the planet (Peale, 1988; Peale *et al.*, 2002; Margot *et al.*, 2007; Margot *et al.*, 2012; Chapter 4). These quantities are the normalized polar moment of inertia C/MR^2 and the fraction of the polar moment of inertia contributed by the portion of the planet that overlies the liquid outer core C_m/C.

The MESSENGER-derived moment of inertia values and the bulk density of the planet have been used to constrain the relative thicknesses of the silicate mantle and metallic core, and their respective densities, in suites of models of varying complexity (Margot *et al.*, 2012; Hauck *et al.*, 2013; Rivoldini and Van Hoolst, 2013; Dumberry and Rivoldini, 2015; Chapter 4). Early estimates for the average density of the outermost solid layer of the planet and the metallic core of 3380 ± 200 kg m^{-3} and 6980 ± 280 kg m^{-3}, where the boundary between these two layers is ~420 ± 30 km below the planet's surface (Hauck *et al.*, 2013), are consistent with the most recent estimates of Mercury's gravity field and rotational parameters (Mazarico *et al.*, 2014) because of the similarity to previous estimates of these parameters (Margot *et al.*, 2012).

Detailed models of Mercury's interior have been designed to resolve additional layers within the interior. For example, as discussed in section 19.3.1, detailed analyses of gravity and topography data returned by MESSENGER have led to improved estimates of the thickness of the silicate crust of 35 ± 18 km or >38 km, depending on the method employed (James *et al.*, 2015; Padovan *et al.*, 2015; Chapter 3). That the average density of the outermost solid shell of the planet is greater than expected for iron-poor silicate materials, together with estimates of the composition of Mercury's core inferred from the strongly chemically reducing conditions discovered at the surface, have led to the consideration of a solid iron sulfide layer at the top of the core (Smith *et al.*, 2012; Hauck *et al.*, 2013; Padovan *et al.*, 2014; Chapter 4). Given that both silicon and sulfur should have partitioned into the core (Chapter 2; Section 19.4.1), at the modest pressures prevalent at the top of the core, melting Fe–S–Si can yield two immiscible liquids (one Fe–S rich and the other Fe–Si rich) over a broad range of bulk compositions. This behavior would lead to segregation of the sulfur-bearing liquids to the shallowest portions of the liquid core, including the core–mantle boundary. Recent metal–silicate partitioning experiments at 100 kPa (1 bar) pressure, however, suggest that the range of potential core sulfur and silicon contents consistent with the surface S content may not lead to core compositions that permit immiscibility and compositional segregation (Chabot *et al.*, 2014) (see also Chapter 4). Additional experimental work at higher pressures and varying silicate compositions are necessary to fully test the importance of liquid immiscibility in Mercury's core and the possibility of a solid FeS layer. However, measurement of induced magnetic fields at Mercury has led to estimates of the depth to the top of the core (Johnson *et al.*, 2016; Chapter 5) that are consistent with internal structure models. Taken together, the consistency between the internal structure models that give an estimate of the depth to the top of the liquid outer core, and the induced magnetic field analyses that yield the depth to the top of an electrically conducting layer, indicates that any FeS layer, if present, is limited in thickness.

Similarly, an Fe-rich solid inner core may also be present, though constraints on its size are sparse. Internal structure models consistent with the gravity field and rotational state of Mercury are generally limited in their ability to resolve the inner core (Chapter 4), though there does appear to be a slight tendency toward relatively modest inner core sizes (Hauck *et al.*, 2013; Dumberry and Rivoldini, 2015), perhaps smaller than half the total core radius. Recent work on the dynamic coupling of the rotation of the inner core to the outer, librating solid shell of the planet indicates that for inner cores larger than ~30% of the radius of the planet, it is necessary to know the size of the inner core in addition to the gravity field and rotation data in order to infer the moments of inertia of the planet (Peale *et al.*, 2016; Chapter 4). Although at present it is not possible to determine independently the size of the inner core, models with inner cores larger than 30% of the radius of the planet tend to have silicate layer densities less than the densities of magnesian olivine and pyroxene, the likely dominant constituents of Mercury’s mantle. Thus, Mercury’s inner core, if present, is unlikely to have a radius more than 30% of the planet’s radius.

19.4.3 Magnetic field

Mercury’s magnetic field observations demonstrate that a global-scale field is presently being generated by a core dynamo (Chapter 5). Initial data from Mariner 10, along with the more recent MESSENGER mission measurements, show that Mercury’s dynamo-generated field is relatively weak and dominated by an axially aligned dipole. The dipole dominance of the field suggests, at first glance, that Mercury’s dynamo may be quite Earth-like in its morphology, although a suite of characteristics of Mercury’s field suggest that it has distinctive properties.

The weak intensity of the field challenges our understanding of how Mercury’s magnetic field is generated. Both energy- and force-balance arguments suggest that Mercury’s observed magnetic field should be at least two orders of magnitude stronger than the field measured by Mariner 10 and MESSENGER. Although the dipole is the largest harmonic in the field, the quadrupole component is relatively large, at approximately 40% of the dipole strength. This quadrupolar component – equivalent to an offset of the dipole from Mercury’s center – is larger than observed for other planets with dipole-dominated fields. Indeed, it is larger than those of other planets even when corrections are made for the relatively shorter distance from the surface to the core–mantle boundary (CMB) at Mercury, with a proportionately smaller attenuation of the quadrupole component with distance from the dynamo region. The multipolar terms beyond the quadrupole, though, are quite small. Furthermore, a property that has not received much attention to date is the axisymmetry of the dipole and quadrupole components. With the possible exception of Saturn, no other planet has a field as axisymmetric as Mercury. The combination of these three characteristics requires alterations to dynamo scenarios previously proposed for Mercury.

The weakness of Mercury's field was the first puzzle to be confronted, and several solutions have been suggested (e.g., Wicht and Heyner, 2014). For example, numerical dynamo models with very large inner cores (Stanley *et al.*, 2005) or with very small inner cores (Heimpel *et al.*, 2005) could produce relatively weak fields. However, current compositional, thermal, and structural models for Mercury's core suggest that the inner core is unlikely to be sufficiently large to satisfy the large inner core models, even if the size of the inner core is weakly constrained at best (see section 19.4.2). Another explanation for the relative weakness of Mercury's field is that the outer portion of the core may be stably stratified, an idea consistent with the small magnitudes of the terms beyond the quadrupole in the field's multipolar expansion. This stratification could be thermal (the result of subadiabatic heat flux at the CMB) or compositional (due to light element segregation) in origin. Such a stably stratified layer may attenuate the field intensity observed at the surface (Christensen, 2006; Christensen and Wicht, 2008), although double-diffusive convection in the stable layer may hinder the attenuation (Manglik *et al.*, 2010). A third suggestion is that feedback between currents generated in Mercury's magnetosphere and those in Mercury's core may result in a weak field state (Glassmeier *et al.*, 2007; Heyner *et al.*, 2011). A fourth possibility is that, if S is the principal light element in the core, temperatures may drop below the melting temperature near the top and the middle of the core in regions often termed Fe-snow zones, where Fe would crystallize and then sink through the core; this situation contrasts with that of Earth, where crystallization first occurs at the center of the planet (Chen et al., 2008). A proposed consequence of such top-down crystallization in Mercury's core is that there could be two separate regions of dynamo generation and that the dipole components oppose each other, yielding a weak net external field (Vilim *et al.*, 2010).

Although these scenarios offer promising avenues for understanding the weakness of Mercury's field, they must also explain the other characteristics of the magnetic field observed by MESSENGER. None of these proposed mechanisms, by themselves, have yet been shown to naturally lead to magnetic fields with large quadrupole components and very axisymmetric fields. The combination of a large quadrupole component and an axisymmetric dipole component is particularly challenging because dynamo theory demonstrates that when a fluid velocity mode excites the generation of the axial quadrupole component, it will also excite the non-axisymmetric dipole component (Bullard and Gellman, 1954). Special circumstances may therefore apply in order to dampen only one of these magnetic modes.

Two recent studies have had some success in this vein. Cao *et al*. (2014) imposed a north–south symmetric thermal perturbation at the CMB in a numerical dynamo model (resulting in higher heat flux at the CMB equator: see Figure 19.3) along with volumetric heat sources throughout the core. Their model matched the dipole–quadrupole dominance and axisymmetry in Mercury observations, but it did not reproduce the relatively low strength of Mercury's field. The likelihood that such a thermal perturbation is present at Mercury's CMB is also unclear. In contrast, a numerical dynamo model by Tian *et al*. (2015) instead imposed a north–south antisymmetric thermal perturbation (i.e., of spherical harmonic degree 1) at the CMB (Figure 19.3), resulting in higher heat flux in the northern hemisphere. In addition, a thin, stably stratified layer was imposed at the top of the core in this model. This combination of properties resulted in a magnetic field that reproduced the dipole–quadrupole dominance, axisymmetry, and the weakness of Mercury's field. The north–south antisymmetric thermal perturbation in this model was justified on the basis of the concentration of smooth plains in Mercury's northern hemisphere (Head *et al.*, 2011).

Recent work by Philpott *et al*. (2014) also suggested that there has been little to no secular variation in the large-scale magnetic field components between the time of the Mariner 10 flybys (1974–1975) and the four years that MESSSENGER was in orbit about Mercury. A study by Stanley and Bloxham (2016) of the Saturnian dynamo suggests that if Mercury possesses a stably stratified layer at the top of the core, and if the magnetic field is very axisymmetric, then very slow secular variation of the field is a natural result. This correspondence between slow secular variation and a stably stratified layer may help to explain the lack of observed secular variation in Mercury's magnetic field.

19.4.4 Core properties

The relative dominance of Mercury's core as a fraction of the planet's mass and volume (Chapter 4) underscores the influence of the core in the planet's overall evolution. The basic properties of the core, and particularly its thermodynamic attributes, are critical for understanding how it has evolved. Siegfried and Solomon (1974) utilized a thermal conduction model for heat transport through the planet, in concert with knowledge of the thermodynamic properties of iron for the core, to investigate the thermal history and core crystallization of Mercury. More recent approaches have generally considered heat transport through the mantle via convection and various alloys of iron and sulfur for the core (e.g., Schubert *et al*., 1988; Hauck *et al*., 2004; Grott *et al*., 2011; Tosi *et al*., 2013).

Over the past two decades, knowledge of the behavior of a variety of potential core-forming materials has grown considerably. The pressures within Mercury's core, ~5–40 GPa (Hauck *et al.*, 2013), are directly accessible in laboratory experiments. Of particular interest are the temperature and pressure dependencies of the properties of iron alloys, including the thermal conductivity, thermal expansivity, and melting behavior. It is well-known that Fe–S alloys have the peculiar behavior that their eutectic melting temperature decreases with increasing pressure (e.g., Fei *et al.*, 1997; Fei *et al.*, 2000; Li *et al.*, 2001; Chudinovskikh and Boehler, 2007; Stewart *et al.*, 2007; Chen *et al.*, 2008) up to 14 GPa with shifts in the eutectic composition toward more Fe-rich compositions at pressures up to at least 40 GPa (Stewart *et al.*, 2007).

Iron–silicon alloys, which may be present in the core as a consequence of Mercury's chemically reduced conditions (see section 19.4.1), behave differently from alloys of iron and sulfur. The primary distinctions are that the presence of silicon results in a smaller melting point depression than with S and the Fe–Si alloys show a strong solid solution (Kuwayama and Hirose, 2004), particularly when compared with the limited solubility of S in solid Fe, even at high pressure (Li *et al.*, 2001). Furthermore, temperature differences between the liquidus and solidus are <50 K and the compositional differences between coexisting liquid and solid are <2 wt % Si on the Fe side of the eutectic in this system at 21 GPa (Kuwayama and Hirose, 2004). In contrast to the Fe–S system, it appears that the eutectic temperature increases with pressure (Kuwayama and Hirose, 2004; Morard *et al.*, 2011; Fischer *et al.*, 2013) up to at least 50 GPa, and the Si content of the eutectic composition increases with pressure up to at least 21 GPa (Kuwayama and Hirose, 2004).

The more likely situation is that Mercury's core contains multiple alloying elements, particularly S and Si (e.g., Chapter 2; section 19.4.1) because of a broad trade-off from S- to Si-bearing Fe alloys as a function of decreasing oxygen fugacity, with mixtures of the two quite likely (Malavergne *et al*., 2010; Hauck *et al*., 2013; Chabot *et al*., 2014). Depending on the composition of the core, liquid immiscibility at pressures less than 12 GPa is possible, with the result that S- and Si-rich liquids would separate due to their differential buoyancy and would lead to a S-rich liquid at the top of the core (section 19.4.2) – though recent experiments suggest that a single miscible Fe–S–Si liquid is more likely (Chabot *et al*., 2014).

In addition to the melting behavior of Fe–rich alloys likely to be present in Mercury's core, there are other thermodynamic properties of these materials critical to the planet's evolution. Among the most relevant is the thermal expansivity of these alloys, which is a controlling parameter in both the temperature gradient and in the amount the planet expands or contracts with temperature changes (Siegfried and Solomon, 1974; Schubert *et al.*, 1988; Hauck *et al.*, 2004; Williams, 2009; Grott *et al.*, 2011; Hauck *et al.*, 2013; Tosi *et al.*, 2013; Jing *et al.*, 2014). Measurement of the thermal expansivity of Fe alloys is challenging, particularly at high pressure and for liquids. The majority of pre-MESSENGER-era models of Mercury's interior were predicated on a constant value for thermal expansivity in the core (e.g., Siegfried and Solomon, 1974; Schubert *et al*., 1988; Hauck *et al*., 2004; Grott *et al*., 2011), consistent with expectations for Earth's core. However, the thermal expansivity is clearly a function of pressure and composition (e.g., Williams, 2009), and recent work on both the internal structure (Smith *et al.*, 2012; Hauck *et al.*, 2013; Rivoldini and Van Hoolst, 2013; Chapter 4) and the contraction of Mercury (Tosi *et al*., 2013) has aimed to accommodate this variation. Recent experimental measurements of the density and sound velocity of Fe alloys at high pressure and temperature have altered our understanding of the variation in core thermal gradients and the potential for contraction in these relevant systems, particularly for Fe–S alloys (Jing *et al.*, 2014). The primary consequence of these new data and models is the potential for steeper adiabatic thermal gradients and larger amounts of thermal contraction than previously appreciated.

The thermal conductivity of Fe alloys is particularly important for understanding both core heat transfer and the evolution of Mercury's magnetic field. Recent experimental and numerical estimates of the thermal conductivity of iron and Fe alloys at the conditions of Earth's core are greater than previous canonical values by a factor of 2–3 (e.g., de Koker *et al.*, 2012; Pozzo *et al.*, 2012; Gomi *et al.*, 2013), leading to questions regarding the relative role of thermal buoyancy in driving Earth's dynamo. However, molecular dynamics calculations by Zhang *et al*. (2015) are more consistent with lower values of the conductivity, as are recent direct measurements of thermal conductivity of solid iron at high pressure (Konôpková *et al*., 2016). At Mercury's core conditions, experimental work also appears to suggest a larger thermal conductivity in pure Fe (Deng *et al.*, 2013b) than is typically found in models of the planet's core heat transport and magnetic field generation. Moreover, Si is known to substantially decrease the thermal conductivity in Fe alloys (Seagle *et al.*, 2013), perhaps reducing even the larger estimates of thermal conductivity toward the values assumed for Mercury's core in previous models of the planet's interior. Ultimately, the thermal conductivity of Mercury's core depends on pressure, temperature, and composition and plays a major role in the thermal gradient and the longevity and pervasiveness of core convection necessary for driving the magnetic field.

19.5 Thermo-chemical models of interior evolution

The volcanic and tectonic evolution of Mercury as recorded on its surface is closely connected to the amount of heat in, and the transfer of that heat from, the planet's interior.

Therefore, understanding Mercury's geologic evolution also requires an understanding of the processes acting in the deep interior. Thermal or thermo-chemical evolution models are usually employed to shed light on the working of a planet's interior heat engine (Solomon, 1977; Stevenson *et al.*, 1983; Schubert *et al.*, 1988; Hauck *et al.*, 2004; Redmond and King, 2007; King, 2008; Grott *et al.*, 2011; Michel *et al.*, 2013; Tosi *et al.*, 2013). In order to understand the evolution of the entire planet we utilize models of planetary evolution that incorporate crucial thermal, chemical, magmatic, and tectonic processes constrained by MESSENGER observations.

19.5.1 Modeling approaches

The most straightforward models of the internal evolution of planets consider the global energy balance for the mantle and core via the relations

$$\rho_{\mathrm{m}} c_{\mathrm{m}} V_{\mathrm{m}} \frac{dT_{\mathrm{m}}}{dt} = -q_{\mathrm{m}} A_{\mathrm{m}} + Q_{\mathrm{m}} V_{\mathrm{m}}$$

$$\rho_{\mathrm{c}} c_{\mathrm{c}} V_{\mathrm{c}} \frac{dT_{\mathrm{c}}}{dt} = -q_{\mathrm{c}} A_{\mathrm{c}} + Q_{\mathrm{c}} V_{\mathrm{c}}$$

where ρ is density, c is the heat capacity, V is volume, T is temperature, A is surface area, and the subscripts m and c refer to the mantle and core, respectively. Q is the rate of heat released in the interior per unit volume by the decay of the long-lived radioactive elements ^{40}K, ^{232}Th, and ^{238}U, and q_{c} and q_{m} are the heat flow out of the core and the mantle, respectively. A model based on the above energy balance is generally sufficient to quantify the amount of secular cooling of the planet, and parameterizations for the heat flow values q_{m} and q_{c} derived from scaling relations between key dimensionless numbers are usually employed to describe the heat transport from the mantle to the surface (e.g., Moresi and Solomatov, 1995; Grasset and Parmentier, 1998; Reese *et al.*, 1998), where heat is ultimately radiated to space.

Perhaps the most important parameter for governing how heat moves through a planetary mantle is the solid-state viscosity. The mantle viscosity has a strong dependence on temperature such that the cool, outermost portion of the planet is rigid, yet at temperatures several hundred to a thousand degrees hotter mantle material behaves like a slow-moving fluid. A consequence of this behavior for most planets is that an immobile upper layer called a stagnant lid rapidly develops. This situation is in contrast to the mechanism of plate tectonics operating on Earth. The slow diffusion of heat from the interior through the thick stagnant lid is considerably less efficient than the advectively dominated heat transport from lithospheric recycling in plate tectonics. As a result, cooling is slow in most planets, and the interior is kept warm over extended periods of time.

Planetary thermal evolution calculated via the parameterized energy balance approach can then be combined with a model of mantle melting behavior to quantify the amount of melt generated in the interior (e.g., Hauck *et al*., 2004, Grott *et al*., 2011). With limited melting experiments tailored to the low-Fe and highly reducing conditions in Mercury's mantle, the well-characterized solidus of terrestrial KLB-1 peridotite has often been used as a proxy for the mantle solidus (e.g., Herzberg *et al.*, 2000). In these one-dimensional parameterized mantle convection models, melt is then generated whenever the mantle temperature exceeds the model solidus and is assumed to be extracted instantaneously, whereas the melt region is replenished with undepleted material on a timescale associated with the mantle convection speed. A general sketch of the relevant temperatures in the interior and the generation of partial melt is shown in Figure 19.4.

One of the major constraints on models of the thermo-chemical evolution of Mercury is how much the planet has radially contracted as documented by its surface tectonic landforms (Chapter 10; Section 19.3.2). Three global processes contribute to radial contraction, and the magnitude of each can be estimated once the thermo-chemical evolution of mantle and core has been calculated (Hauck *et al*. 2004; Grott *et al*., 2011; Tosi *et al*. 2013). Cooling causes the mantle and core to contract, resulting in a contribution ΔR_{th} to the change in planetary radius. Phase changes in the core and mantle can result in changes in the specific volume of their constituent materials, which further contribute to a change in the radius of the planet. Usually, consideration of phase changes is restricted to partial melting of the mantle and freezing of an inner core. The products (i.e., crust and the mantle residuum) of mantle differentiation have a larger volume than the primordial mantle, resulting in a net expansion, ΔR_{md}, of the planet (Kirk and Stevenson, 1989). However, solidification of the solid inner core results in a decrease in volume and a hence a radial contraction, ΔR_{ic} (Solomon, 1976). However, with Si present in the core, the density difference between the liquid and solid phase will be small as a result of the nearly similar compositions of the two phases. In total, the radius change of the planet can be expressed as the sum of the individual contributions

$$\Delta R_{\mathrm{p}} = \Delta R_{\mathrm{th}} + \Delta R_{\mathrm{md}} + \Delta R_{\mathrm{ic}}$$

where the details of the amount of radius change depend on the assumed chemical composition of mantle and core as well as on the associated thermal expansion coefficients (e.g., Grott *et al.*, 2011; Tosi *et al.*, 2013).

Whereas one-dimensional models have been shown to be sufficient to study the global evolution in terms of secular cooling, crustal production, and planetary contraction (Tosi *et al.*, 2013), more complex two- and three-dimensional models are necessary to understand the planform of mantle convection, the efficiency of mantle mixing, and the persistence of mantle convection to the present (Redmond and King, 2007; King, 2008; Michel *et al.*, 2013; Tosi *et al.*, 2013). Instead of parameterizing the heat flow into and out of the mantle, such models involve self-consistent solutions to the equations of mass, energy, and momentum transport in the mantle and directly calculate convective velocities and the temperature distribution in the interior. The chemical composition of the mantle is often tracked with particle tracers (Plesa *et al.*, 2013; Tosi *et al.*, 2013), and the resulting buoyancy is included in the momentum conservation equation.

The increase in model detail of two- and three-dimensional simulations comes at the price of higher computational cost, and running a large number of Monte-Carlo style simulations, as is increasingly common with one-dimensional models, becomes prohibitively expensive for fully dynamical models. Instead, the parameter space is usually sampled with a few representative models. Depending on the aim of the investigation, model complexity can be reduced by considering two-dimensional models (Redmond and King, 2007; Michel *et al*., 2013), or by disregarding crustal production (Redmond and King, 2007; King, 2008) or mantle mixing of melt residuum (Redmond and King, 2007; King, 2008; Michel *et al*., 2013).

The other major constraint on the internal evolution of Mercury is its internally generated magnetic field. A magnetic field generated by a core dynamo requires fluid motions within the electrically conductive fluid portion of the core. A commonly employed minimum, though not necessarily sufficient, requirement for dynamo generation is that if the motions are the result of thermal convection then the core heat flux must exceed the amount of heat that can be transported by thermal conduction along the adiabatic thermal gradient that convection imparts. Energy for driving convective motions also may be derived from compositional buoyancy, such as is generated by the expulsion of a relatively light element rich fluid upon the crystallization of the core. It has also been suggested (e.g., Christensen, 2006) that Mercury's core may not be entirely convecting and could instead have a stable layer at its top that may account for its relatively weak magnetic field (see section 19.4.3). In both of these latter cases, the heat flux at the CMB flows from the core to the mantle but may be less than can be conducted along the adiabatic thermal gradient, with convection restricted to deeper portions of the core.

19.5.2 Persistence of mantle convection

Mercury's large core and relatively thin silicate shell raise important questions about how the planet has cooled through its history, in particular the role of mantle convection. These questions are important because upwelling mantle is generally a critical ingredient in magma generation, and convection leads to larger rates of cooling that help drive the fluid flow in the core necessary for magnetic field generation. Solid-state convection within a layer in a planetary body depends on several material properties as well as the temperature contrast across the layer and depends strongly on the thickness of that layer. For bottom-heated convection, the vigor of mantle convection is described by the non-dimensional mantle Rayleigh number

$$Ra = \frac{\rho g \alpha \Delta T D^3}{\kappa \eta}$$

where ρ is density, g is gravitational acceleration, α is thermal expansivity, ΔT is the temperature difference across the convecting layer, κ is thermal diffusivity, η is mantle viscosity, and $D = R_p - R_c$ is the thickness of the convecting layer, where R_p is the planetary radius and R_c is the core radius. Convection requires that Ra be larger than some critical value, and above that value the vigor of convection increases with Ra. Therefore, mantle convection is more difficult in a thin mantle than in a thicker one because of the cubic dependence on layer thickness. As a consequence, mantle convection in Mercury is expected to be sluggish compared with convection in planets with thicker mantles. Furthermore, mantle cooling can result in a transition from convection to thermal conduction in the mantle if the Rayleigh number falls below its critical value. The cessation of mantle convection would result in the end of pressure-release melting during convective ascent and most likely the end of global-scale volcanism, though local volcanism could continue, such as observed in impact basins and sites of small-scale explosive volcanism (Chapter 11).

As a result of Mercury's thin mantle, predictions of rather modest internal heat production, and the strong temperature-dependence of the viscosity of mantle rocks, questions were raised in the pre-MESSENGER era about the persistence of mantle convection to the present. Some studies with one-dimensional parameterized convection models found that, although convection was important for much of the planet's history, it may have ceased before the present (Hauck *et al.*, 2004). However, other work with two- and three-dimensional fluid dynamic models (Breuer *et al.*, 2007; Redmond and King, 2007), as well as studies with one-dimensional models that considered the insulating capacity of the near-surface regolith and crust (Grott *et al.*, 2011), generally found that mantle convection persisted throughout the planet's history.

MESSENGER's observations of Mercury have substantially improved our understanding of the planet's interior and so have helped refine many of the assumptions and boundary conditions required for models of its internal evolution. The most important of these constraints are the improved knowledge of radiogenic heat production (Peplowski *et al.*, 2011, 2012) and the thickness of the outer solid shell of the planet (Chapter 4). Typically, earlier work was based on the assumption that the core–mantle boundary was ~600 km deep compared with the ~420 km determined by MESSENGER (Hauck *et al.*, 2013).

That data from MESSENGER indicate Mercury's silicate shell is nearly one-third thinner than in previous models has led to a reevaluation of whether mantle convection continued throughout the planet's history. Furthermore, although the precise partitioning of heat-producing elements between the near surface (where they have been measured) and the interior is only weakly constrained, the relative amounts of U, Th, and K, as well as their surface abundances, provide important (and previously unavailable) constraints. Indeed, the finding of surprisingly abundant K (Peplowski *et al.*, 2011, 2012) is important for quantifying Mercury's internal evolution because of the strong heat output of ^{40}K coupled with its relatively shorter half-life than the long-lived isotopes of U and Th. Taking these new data into account, Michel *et al*. (2013) reevaluated the issue of convection within Mercury's mantle utilizing two-dimensional axisymmetric, spherical shell fluid dynamic calculations. They found that, for a broad range of conditions of mantle heat production, mantle viscosity, and initial internal temperatures, cessation of mantle convection within the past several billion years is common in models with silicate layers less than ~440 km thick. These results are consistent with those of Tosi *et al*. (2013), who evaluated the internal evolution of Mercury in one-, two-, and three-dimensional models of mantle convection additionally constrained by ~3 km of global radial contraction as had been inferred from mapping ~21% of Mercury's surface by Di Achille *et al*. (2012). In the models of Tosi *et al.* (2013), cessation of mantle convection within the past 1–1.5 Gyr was the norm. However, the 5–7 km of radial contraction inferred from more recent global mapping (Byrne *et al.*, 2014) warrants additional thermal evolution calculations, because a larger total cumulative contraction may require higher rates of cooling that may be more consistent with prolonged mantle convection than with thermal conduction only. As a result of its small obliquity, its proximity to the Sun, and its 3:2 spin–orbit resonance (Chapter 4), Mercury has large spatial variations in surface temperature. By including the latitudinal variation in temperature, Michel *et al*. (2013) found that cessation of mantle convection may be delayed by a few hundred million years relative to typical models with a spatially constant surface temperature.

With a Monte-Carlo approach and the inferred magmatic evolution and global contraction as model constraints, Tosi *et al*. (2013) determined the times at which convection stopped in one-dimensional models of Mercury's thermo-chemical evolution. An update of their calculations, taking into account the larger amount of global contraction and the observation that Mercury had an ancient magnetic field as well as a modern one, is shown in Figure 19.5. In all, about 40% of the models consistent with the presently available constraints are found to convect to the present. This outcome is a direct consequence of the more recent estimate of global contraction, which allows for Mercury to have experienced more efficient mantle cooling than in the models of Tosi *et al*. (2013), which permitted only 3 km of contraction and had vanishingly few outcomes in which mantle convection operated at present.

19.5.3 Internal evolution models consistent with observational constraints

Our current understanding of the timing of major processes in Mercury's evolution, as described in the preceding sections, is summarized in Figure 19.6. Evidence from the first ~500 Myr is limited mainly as a result of resurfacing by intercrater plains formation and impact cratering before and during the late heavy bombardment. An internally generated magnetic field was active prior to 3.7–3.9 Ga (Johnson *et al.*, 2015; Chapter 5) and is active today (Ness *et al.*, 1976; Ness, 1979; Anderson *et al.*, 2011, 2012), implying a cooling core within which either thermal or chemical convection operated during each era. The magnetic field history between ~3.8 Ga and the present is presently unknown, and either a continuously operating core dynamo or an early shutdown of the dynamo followed by a later reinitialization are plausible scenarios

(Chapter 5). Effusive volcanism was widespread early in Mercury's recorded history (e.g., Marchi *et al.*, 2013), and the areal extent of volcanism waned rapidly from the LHB until perhaps ~3.5 Ga, after which effusive activity largely ended, with the exception of local activity within younger impact basins (Prockter *et al.*, 2010; Denevi *et al.*, 2013a; Byrne *et al.*, 2016; Chapter 11). Explosive volcanism continued for a longer time period than did widespread plains volcanism (Kerber *et al.*, 2009; Thomas *et al.*, 2014). The global contraction accumulated on shortening tectonic landforms records planetary cooling from the end of the LHB to the present (Chapter 10).

Following the approach of Tosi *et al*. (2013), the range of models that satisfy the following major constraints can be determined. Successful models must (1) produce at least 5 km of crust by partial melting of the mantle, which is a minimal requirement for producing the intercrater and smooth plains, (2) show 5 to 7 km of global contraction following the end of the late heavy bombardment, and (3) exhibit heat flow from the core that would permit, though not require, the generation of a magnetic field. It is worth noting that the choice of the thickness of extracted crust has little influence on the results, as long as some crust is produced. Furthermore, the requirement on heat flow from the core serves to reject those models that would preclude a thermally driven core dynamo during the earliest evolution, but is not particularly restrictive later in the planet's history as core heat flux is generally small after 4 Gyr of evolution.

Models that satisfy all of these constraints show some common trends. Slow cooling of the planet is required, and model mantle reference viscosities at 1600 K range from 10^{20} to 10^{22} Pa s. Additionally, most models also show an early phase of mantle heating, whereas the core cools monotonically throughout evolution. Also, although up to 100 km of crust can be produced, most models produce less than 75 km. Furthermore, surface heat flow declines from about 30 mW/m^2

at the beginning of evolution to ~10 mW/m^2 today, consistent with an estimate derived from tectonic modeling (Egea-González *et al.*, 2012). Finally, and most interestingly, the ratio of the concentration of heat-producing elements in the crust to that in the primordial mantle is found to be between 2 and 4.5, which is similar to the results obtained by Tosi *et al*. (2013) for a core radius of 1940 km. On the other hand, the initial mantle temperature in the models is poorly constrained and can range from 1600 to 1900 K, similar to the range in initial core temperature.

A typical thermo-chemical evolution model that satisfies the above constraints is shown in Figure 19.7, where the core and mantle temperature; the core, mantle, and surface heat flow; the radius change from thermal contraction and mantle differentiation; and the crustal thickness, stagnant lid thickness, and extent of the partial melt zone are shown as functions of time. In this model, the initial mantle temperature is 1700 K, the initial core temperature is 1875 K, the crustal thermal conductivity is 2.5 W m^{-1} K^{-1}, a poorly conducting regolith layer 5 km thick and with a thermal conductivity of 0.2 W m^{-1} K^{-1} is included, and the mantle viscosity is $10^{20.5}$ Pa s. With surface abundances of radiogenic elements of 1288 ppm ^{40}K, 155 ppb ^{232}Th, and 90 ppb ^{238}U (Peplowski *et al.*, 2012) and a crustal enrichment factor of 3.5, this typical model has bulk silicate concentrations of heat-producing elements of 368 ppm ^{40}K, 44 ppb ^{232}Th, and 25 ppb ^{238}U, similar to values for Earth and Mars. Following the late heavy bombardment (i.e., ~3.8 Ga), the model monotonically cools at a rate of 40 K Gyr^{-1}, with the core and mantle cooling at the same rate. Global crustal production ceases around 2.5 Ga (though is largely complete nearly 1 Gyr earlier), and a total of 25 km of crust is produced, resulting in a final crustal thickness of 30 km. Total radial contraction is just short of 7 km, with continuous accumulation of contraction following the late heavy bombardment. It is worth noting that care must be taken when interpreting the timing of crustal production from such one-dimensional models, as this timing can differ considerably from that determined with fully dynamical two- or three-dimensional models, which generally have crustal production concentrated earlier in the planet's evolution but result in similar total crustal thickness values (e.g., Tosi *et al*., 2013).

Given the uncertainties associated with the state and composition of Mercury's core, the model shown in Figure 19.7 focuses on the most robust aspects of the core and considers only thermal contraction of the core and does not take into account contraction by core solidification. Although, for a given amount of inner core growth, this solidification could be a major contribution to planetary contraction for an Fe–FeS core composition (e.g., Solomon, 1976; Schubert *et al.*, 1988; Knibbe and van Westrenen, 2015), it would be less so if Si were the major alloying light element in the core (Fei *et al.*, 2011) as the density difference between solid and liquid would be smaller because of the very small difference in Si content between solid and liquid (Kuwayama and Hirose, 2004). However, the melting behavior of core material is an important factor in core contraction arising from crystallization: S-bearing cores would experience less inner core growth due to the stronger melting point depression relative to Si-bearing alloys. The true contribution of core freezing to global contraction will likely fall between these two limiting cases, but this effect is difficult to quantify without further data on the equation of state of the Fe–S–Si system. More importantly, it is clear that there is little room for a large contribution to the observed global contraction from core crystallization. The solidification of a large volume fraction of the core would lead to significantly more total contraction than that from thermal contraction alone, e.g., crystallization of > 2.5% the volume of the core (equivalent to an inner core <30% of the radius of the core) would lead to at least 2 km of additional contraction (Grott et al., 2011). Thus, the contribution of core crystallization is likely limited, as fewer models would be permitted because they would exceed the 7 km of radial contraction accommodated by tectonic deformation and even the 9 km inferred for total planetary contraction that includes the elastic accommodation of radial contraction prior to the formation of major faults (Chapter 10). This result implies that either core solidification was close to complete by the end of the late heavy bombardment, or that only a small inner core started freezing in the recent past. Because of indications that the inner core is likely small (Chapter 4), the latter scenario is more likely.

A three-dimensional view of the thermal evolution of a model with the same properties as discussed above is shown in Figure 19.8. Additionally, the surface temperature variation imposed by Mercury's 3:2 spin–orbit resonance is taken into account (Chapter 4). The model is similar to that presented by Tosi *et al*. (2015), in which chemical composition is tracked with a particle tracer technique (Plesa *et al.*, 2013), and uses the same initial conditions as the model shown in Figure 19.7. Figure 19.8a shows the variation of average annual surface temperature, which ranges from 260 to 430 K between the poles and the equatorial regions. The mantle convection pattern shown in Figure 19.8b reflects this type of temperature distribution, with downwellings (blue) more focused near the polar regions. As a result of the small thickness of Mercury's mantle, the convective pattern shows only small-scale up- and downwellings, and the more linear structures found in earlier simulations of mantle convection with a mantle thickness of 600 km (King, 2008) are not reproduced. Toward the end of the model run, mantle convection ceases, resulting in a conductive temperature profile in the mantle (Figure 19.8c). In this model, modern mantle temperatures reflect the forcing imposed by the insolation pattern. However, it should be noted that it takes a few hundred million years for the perturbation from insolation to diffuse to any meaningful depth. Therefore, the full extent of the temperature forcing will be reflected in the deep interior only if the 3:2 spin–orbit resonance has been stable for an extended period of time (Correia and Laskar, 2004; Noyelles *et al.*, 2014).

Although the general picture of Mercury's thermo-chemical evolution is consistent with the constraints provided by MESSENGER observations, details of the models may change as more data are analyzed and eventually provided by new missions such as BepiColombo (Chapter 20). In particular, the amount of radial contraction documented in shortening tectonic structures has been continuously refined (Strom *et al.*, 1975; Watters *et al.*, 2009; Di Achille *et al.*, 2012; Byrne *et al.*, 2014), resulting in less stringent constraints on Mercury's thermal evolution. Current best estimates for the total radial contraction accumulated by brittle structures since the late heavy bombardment range from 5 to 7 km (Byrne *et al.*, 2014) but may be as large as ~9 km when elastic deformation is considered, or less than 5 km if the dip angles of the thrust faults are uniformly and surprisingly steep (Chapter 10). Importantly, larger values (> 7 km) of contraction would allow for lower mantle viscosities and thus more efficient mantle convection. Alternatively, such greater contraction could also allow for a larger contribution of core solidification to the total contraction of Mercury, depending on core composition, or more likely some combination of increased cooling and core solidification.

19.5.4 Other factors influencing Mercury's thermo-chemical evolution

One of the factors not considered in the above models is the potential presence of heat-producing elements in Mercury's core. At the low oxygen fugacities inferred from the high S abundance and low FeO content in Mercury's crust (Zolotov *et al.*, 2013), lithophile elements such as K, Th, and U can become more siderophile (Malavergne et al., 2010). McCubbin *et al.* (2012) estimated that up to 10% of the total inventory of U and potentially Th could have partitioned into the core, thus providing an additional heat source that could slow global contraction. However, the differences in global contraction between models with and without heat-producing elements in the core have been found to be minor (Tosi *et al*., 2013), as the total inventory of heat-producing elements in the interior is only weakly affected. Partitioning of U and Th into the core tends to increase the heat flux out of the core and can extend the period during which a thermal-buoyancy-generated dynamo can operate by as much as 100 Myr.

In addition to the production of partial melt in the interior, Mercury's surface compressive stress state has likely been an important factor controlling effusive volcanism. On a contracting planet such as Mercury, extrusive volcanism may be substantially inhibited as magma pathways to the surface are shut-off by maximum compressive stresses in the horizontal direction (Chapter 11). Therefore, the longevity of volcanism as observed on the surface may not be a direct indicator of the timing of melt production in the deep interior. On the other hand, local factors such as variations in the thickness of an insulating crust and/or regolith layer, which would have a lower thermal conductivity than the mantle (Section 19.5.3), largely due to higher porosities (Schumacher and Breuer, 2006), are usually not fully taken into account in thermo-chemical evolution models. Therefore, local volcanism may be ongoing even if global models, particularly one-dimensional models, do not predict the production of partial melt at a given time.

Another energy source not treated in the above discussion is impact heating, which would be expected to contribute to the global energy balance mainly during the early phases of Mercury's evolution. Impact heating associated with the formation of the Caloris impact basin has been modeled by Roberts and Barnouin (2012), who showed that impact heat can alter mantle dynamics. In addition to the production of melt at the impact site itself, partial melting may be induced even far from the impact. Thus, the smooth plains within and adjacent to the Caloris basin could be at least in part the consequence of the impact itself, the heat for which is stored in the mantle over an extended period of time. On the other hand, the influence of isolated impacts on the global evolution of the planet is relatively small (Roberts and Barnouin, 2012), and the conclusions drawn from the simpler models discussed above remain essentially unchanged.

19.5.5 Core evolution

MESSENGER's unveiling of Mercury's internal structure and the geometry and history of its internal magnetic field underscore the important role of the metallic core on the planet's evolution. Taken in concert with the growing understanding of the properties of materials at the conditions of Mercury's core (section 19.4.4), which indicate the potential importance of zones of top-down crystallization and liquid–liquid immiscibility, it is clear that core evolution in Mercury differed from that of Earth's core. Ultimately, models of core evolution on Mercury must account for the planet's magnetic field structure and history (Chapter 5), match the internal structure (Chapter 4), and be consistent with the magnitude of the planet's contraction (Chapter 10).

The driving mechanisms of core evolution are cooling and the chemical differentiation that results from crystallization as the core cools below its melting temperature. The rate of core cooling depends strongly on how the mantle is cooling, as all of the heat from the core must pass through the mantle on its way to the planet's surface. Early in the planet's history, core cooling may have been relatively rapid (Figure 19.7), especially if the planet was hot, because high internal temperatures reduce the viscosity of the mantle and make it easier to remove heat quickly by convection. Of course, just as the cooling of the mantle slows as its initial store of heat of formation is lost and heat production follows the decay of radioactive elements, the cooling of the core slows as well. The rate of cooling of the core is important because a source of convection is necessary to drive the motions in the electrically conductive liquid metal that generate the magnetic field. A minimum condition for thermal convection throughout the entire core is that the heat flux through the CMB must exceed that which can be conducted along the adiabat. Given a thermal conductivity of 40 W m^{-1} K^{-1}, previous workers (Hauck *et al.*, 2004; Tosi *et al.*, 2013) found the minimum core heat flux for thermal convection to be in the range of 12–19 mW m^{-2} for a range of possible thermal expansivity values. Such core heat fluxes are exceeded only early in Mercury's history. The more recent, higher estimates of the thermal conductivity of pure iron at pressures near that of Mercury's CMB (Deng *et al.*, 2013b) of 40–120 W m^{-1} K^{-1} could increase this minimum heat flux by up to a factor of 3. Such high thermal conductivities would limit thermally driven core convection to a very short time period following planet formation. However, the presence of light alloying elements tends to decrease the thermal conductivity; for example, as little as 9 wt% Si reduces the thermal conductivity of the Fe alloy to 41–60 W m^{-1} K^{-1} (Seagle *et al.*, 2013) at Earth's core conditions. As Mercury's core likely hosts considerable abundances of light elements (section 19.4.1), the earlier value adopted for thermal conductivity may not be far off, though the uncertainty may be considerable.

Although it is possible that Mercury's early magnetic field (Chapter 5) was driven by thermal convection, the present-day field is likely dominated by flows driven by compositional buoyancy. The simplest mechanism for generating compositional buoyancy is crystallization of a core alloy in a situation where the compositional difference between the precipitating solid and residual liquid is large, such as has been previously described in the Fe–S system. Sulfur-bearing systems are the best-studied analog for Mercury because of the broad literature on Fe–S melting and because sulfur has such a large melting point depression even at high pressure (e.g., Fei *et al.*, 1997). The consequence of the decreasing melting and eutectic sulfur contents with increasing pressure (section 19.4.4) is that, if the core is comprised of an Fe–S alloy, then it is likely that the crystallization of core material at these pressures began at the top, rather than the bottom, of the core (Hauck *et al.*, 2006; Stewart *et al.*, 2007; Chen *et al.*, 2008; Williams, 2009). An interesting consequence of the combination of the shifts in eutectic temperature and compositions, which vary with pressure, is that two radially separated regions of the core may experience such top-down crystallization, also termed Fe snow (Chen *et al*., 2008).

Both pre- and post-MESSENGER models (Chen *et al.*, 2008; Dumberry and Rivoldini, 2015) of an Fe–S core indicate multiple modes of crystallization, including bottom-up (like Earth) and top-down (Fe snow). In such a system, at low sulfur contents of ~5 wt % or less and with small inner cores, Dumberry and Rivoldini (2015) found that bottom-up crystallization would be expected. However, those workers did not model the non-ideal mixing behavior observed at 14 GPa in the Fe–S system (Chen *et al.*, 2008), which essentially requires a zone of Fe snow between 10 and 14 GPa at even very small sulfur contents because the decrease in melting temperature is so large. With larger sulfur contents or with larger inner core sizes, various top-down crystallization regimes are possible, whether there is a layer of crystallizing material overlying a layer in which the Fe snow re-melts, whether the crystallizing material simply falls to the top of the growing inner core (Hauck *et al.*, 2006; Dumberry and Rivoldini, 2015), or even if there is a second layer of top-down crystallization (Chen *et al.*, 2008).

Top-down crystallization is a consequence of a situation in which the melting temperature increases as a function of depth more slowly than the actual temperature (Hauck *et al.*, 2006; Williams, 2009). In the Fe–S system there is a marked decrease in the eutectic melting temperature with increasing pressure, as well as a reduction in the S content of the eutectic with increasing pressure, both of which lead to melting temperatures decreasing with depth for a wide range of bulk compositions. Measurements of the density and sound velocity of Fe–S liquids at high pressure also indicate that S tends to result in larger adiabatic temperature gradients relative to pure Fe liquids, enhancing this effect and extending to even lower sulfur contents (Jing *et al.*, 2014). As a result of the small melting point reduction in Fe–S alloy cores with low abundances of S, such systems tend to have large inner cores, which in turn tends to concentrate S in the outer core because of the low solubility of S in solid Fe. As a consequence, Fe–S-dominated cores are likely to have experienced Fe-snow regardless of their composition. However, such large inner cores are not favored in structural models constrained by Mercury's rotational dynamics (Chapter 4).

Even though our understanding of the evolutionary paths of Mercury's core under scenarios in which sulfur is the sole light element is becoming more mature, it is also clear that other light elements in addition to, or instead of, sulfur are likely present in the core (Section 19.4.1; Chapter 2). As noted above, carbon is generally a siderophile element, but it has been suggested that C is present as graphite in the mantle and that graphite may have formed an early floatation crust on the planet (Vander Kaaden and McCubbin, 2015), an idea that is consistent with spectral reflectance and neutron spectroscopy observations of the surface (Murchie *et al.*, 2015; Peplowski *et al.*, 2015a, 2016). Consequently, if the core and mantle formed in equilibrium then the core may be saturated in C, although the total amount would be small as the maximum solubility of C in Fe is ~4 wt % and that value decreases with increasing pressure (Lord *et al.*, 2009). This value would be larger if Fe_3C were present, but the density and compressibility of C-bearing alloys are such that it would be difficult for C to be the sole light element in Mercury's core. However, the consequences of even some C being present might be important. For example, the decreasing amount of C in eutectic melts with increasing pressure in the $Fe–Fe_3C$ system is consistent with top-down crystallization, even without S.

In contrast, the presence of silicon, which is likely because of the planet's strongly reducing conditions (see sections 19.4.1 and 19.4.4), has rather different implications for the evolution of the core. The melting behavior of Si-bearing Fe alloys at conditions appropriate to Mercury is more poorly known than for alloys with S or even C. The phase diagram of Fe–FeSi at 21 GPa determined experimentally by Kuwayama and Hirose (2004) is instructive, as they found that the eutectic point is at both a higher temperature and a larger Si abundance than at 100 kPa (1 bar). They also found, as noted above, that the difference in composition between the coexisting solid and liquid phases at temperatures between the solidus and liquidus on the Fe side of the eutectic is very small: a maximum of ~2 wt % Si between the solid and liquid phases. An important consequence of this behavior is an Earth-like bottom-up crystallization of the core, but with residual liquids left by crystallizing of Fe–Si core material that would be only marginally less dense than surrounding material, limiting the buoyancy available to drive convection were the core sufficiently chemically reduced that silicon were the only light alloying element present.

Perhaps most critical to understanding the evolution of Mercury's core is the behavior of Fe alloys with combinations of S, Si, and possibly C. Despite the fact that the thermodynamic properties of multi-component Fe alloys are less well known than for the binary systems, the data that are available suggest interesting evolutionary paths for Mercury's core. For example, liquid immiscibility, such as displayed in both Fe–S–C (e.g., Dasgupta *et al.*, 2009) and Fe–S–Si liquids (section 19.4.1), has potential consequences for compositional segregation within the outer core. Fe–S–C immiscibility would have an influence within only a relatively thin layer near Mercury's CMB because immiscible behavior occurs only at pressures less than 6 GPa

(Dasgupta *et al.*, 2009), which is close to the possible CMB pressure (Chapter 4). However, immiscibility in the Fe–S–Si system would extend deeper within Mercury's outer core (section 19.4.4). Such segregation, if present, likely developed early in the planet's history during metal–silicate differentiation and immediately thereafter. For bulk core compositions near the miscibility limit, however, there is a possibility that the progressive crystallization of a Fe–Si-rich solid and resultant increase in concentration of S in the liquid would drive Mercury's core into a liquid immiscibility state later in its evolution. For this situation to occur, however, relatively large inner core growth would be required to substantially change the outer core composition, an outcome that is inconsistent with models of Mercury's thermal contraction discussed above and estimates of the planet's internal structure (Chapter 4).

A relative lack of experimental data limits firm conclusions about the crystallization behavior in an Fe–S–Si core. Recent experimental results on the Fe–S–Si–C system (Martin *et al.*, 2015) indicate eutectic melting temperatures similar to those of the Fe–S–C system from ~4–15 GPa, with minimal pressure-dependence of the eutectic. Top-down crystallization would be favored in that system. However, data on the pressure-dependence of melting in the Fe–S–Si system are not available at present. While the melting behavior of the Fe–S and Fe–S–Si–C systems suggest that top-down crystallization is likely, the Fe–Si system appears more consistent with a bottom-up crystallization sequence; whether the effects of alloying with S or Si would dominate that behavior is unclear without further data. Determination of melting behavior in the Fe–S–Si system, and the thermodynamic properties that control the adiabatic temperature gradient, are crucially needed in order to understand more fully the crystallization of Mercury's core.

19.6 Discussion

MESSENGER observations have substantially altered our understanding of how Mercury has evolved over its history, but several crucial questions remain open. In particular, we are at a relatively early stage in understanding the connection between the dynamics of the mantle and the production of the crust and the generation of the magnetic field. We next discuss these issues in more detail, focusing on open questions that may be addressed through a combination of analysis of MESSENGER data, modeling, and the acquisition of new observations from BepiColombo and other future missions.

19.6.1 Crustal production and mantle dynamics

Global crustal production through time is a primary indicator of the evolution of a planet – that of its crust and of the interior from which the crust was derived. For planets without crustal recycling, the crust represents a nearly complete time history of intrusive and extrusive volcanism. This history, even when known only to first order, places powerful constraints on our understanding of the evolution of the interior (e.g., Hauck and Phillips, 2002). On Mercury, the clearest constraints on crustal formation are that the best estimate of its average thickness is approximately 35 km (James *et al.*, 2015; Padovan *et al.*, 2015; Chapter 3) and that the tail end of the era of effusive volcanism postdates the Caloris impact by perhaps a few hundred million years at most (Byrne *et al.*, 2016; Chapters 6, 11). Intercrater plains, also interpreted to be dominantly volcanic in origin, are more areally extensive than the smooth plains and are in places as old as 4.0–4.1 Ga (Chapter 6). The first ~500 Myr of Mercury's surface history is also veiled by the overprinting of the late heavy bombardment. Regardless, MESSENGER observations have revealed that Mercury has experienced the most efficient and complete differentiation of mantle and crust among the terrestrial planets, and that this crust was largely built up by successive episodes of effusive volcanism that were likely largely complete within the first 1 Gyr of planet history. Given that Mercury has such a thin mantle, prone to relatively sluggish mantle flow and even the cessation of mantle convection entirely, it is remarkable that generation of the crust could have been so efficient and rapid – particularly in light of the idea that crustal products of a magma ocean may have been only meters thick (e.g., Vander Kaaden and McCubbin, 2015), leaving virtually all of the crust to be produced by serial magmatism. However, because of the low FeO content and modest pressures in Mercury's mantle, the partial melts produced throughout the mantle would be exceptionally buoyant and less susceptible to stalling during ascent (Vander Kaaden and McCubbin, 2015), perhaps facilitating such efficient crustal formation.

The heterogeneity of Mercury's crustal production as observed in its geochemical diversity (e.g., Weider *et al.*, 2015; Chapters 2, 7), and the spatial distribution of smooth plains volcanism, also provide important clues to the history and dynamics of the interior. Indeed, observations by MESSENGER's suite of geochemical sensors indicate both a range of crustal compositions that point to partial melting from multiple sources (Charlier *et al.*, 2013), and a spatial heterogeneity in compositions that does not always follow geomorphological unit boundaries (Peplowski *et al.*, 2015b; Weider *et al.*, 2015). Interestingly, in a manner similar to the Moon's spatial dichotomy in mare volcanism between its near and far sides, and the asymmetric concentration of volcanism on Mars near the Tharsis province, there is a distinctive spatial difference in the abundance of smooth plains units between Mercury's northern and southern hemispheres (Chapters 6, 11). The largest expanses of smooth plains reside at high northern latitudes and within and around the Caloris basin, which is also located in the northern hemisphere. Smaller areas of smooth plains are found generally in proximity to impact basins with little difference in areal coverage between the hemispheres (Chapter 6). Consequently, the processes responsible for the formation of smooth plains in the Caloris region and the northern volcanic plains may be different from those that yielded the isolated, small smooth plains units distributed more evenly throughout the northern and southern hemispheres. Any hemispherical differences in the earlier volcanic activity that produced the intercrater plains are not clear at this time, though some regions also appear to be associated with impact basins (Denevi *et al.*, 2013b). Although MESSENGER provided global geochemical coverage of Mercury, the spacecraft's highly eccentric orbit and high northern periapsis resulted in measurements only at low spatial resolution in the southern hemisphere. That these measurements cannot resolve distinct geochemical terranes in the southern hemisphere limits our understanding of the global evolution of Mercury. The planned orbit for the Mercury Planetary Orbiter on the BepiColombo mission (Chapter 20) will yield higher-resolution southern hemisphere data and may resolve additional geochemical terranes analogous to those observed by MESSENGER in the northern hemisphere.

These heterogeneities in the geochemical and volcanic character of the surface are largely connected to the thermal and chemical properties of the mantle. Mercury's thin mantle yielded a style of mantle convection that was both relatively sluggish and strongly spatially restricted, because the size of individual convective cells would have been on the order of the thickness of the mantle itself. Thus, the large expanses of volcanism in the northern hemisphere require conditions that either permit extraordinarily voluminous magma production from spatially restricted upwellings or conditions that substantially altered the mantle flow dynamics from that expected on the basis of Mercury's mantle thickness. One such mechanism for altering those dynamics is a large impact, such as that which formed the Caloris basin. Indeed, the large thermal perturbation imparted by shock heating from the Caloris impact event may have led to substantial heating of the shallow mantle beneath the impact, but it might also have enhanced some nearby, pre-existing mantle upwellings that generated magma distal from the impact site (Roberts and Barnouin, 2012). Such a mechanism could have been a major contributor to the generation of the Caloris-centric volcanism, but the northern volcanic plains do not appear to host such a large impact capable of triggering such volcanism, even though Caloris and the northern volcanic plains have indistinguishable crater size–frequency distributions and thus ages (e.g., Ostrach *et al.*, 2015). On the other hand, both the broad geochemical heterogeneity across the surface and the smaller, more distributed areas of smooth plains on Mercury could be direct consequences of the small, spatially restricted upwellings and inefficient mixing in a mantle of small thickness. This fluid dynamic behavior of the mantle could act to preserve large-scale geochemical heterogeneities, yet also focus volcanism in locally restricted areas. An important question regarding the era of dwindling effusive volcanism is the relative importance of the pattern of convection (e.g., small yet relatively abundant upwellings) to the total cooling of Mercury that led to a strongly compressive stress state, one that tended to favor intrusive over extrusive volcanic activity.

19.6.2 Evolution of the core and magnetic field

The operation of an internally generated planetary magnetic field is a fundamental indicator of the dynamical behavior of the deep interior of a planet. MESSENGER observations of Mercury's magnetic field have provided important constraints on the character of field generation at present as well as early in the planet's history. Orbital measurements of the geochemical character of the surface materials, as well as gravity and rotational state determinations by MESSENGER, also help characterize the core. However, these new observations raise a number of interesting questions about the behavior of the interior over the history of the planet. In particular, the mechanism of magnetic field generation may require a number of special conditions in order to produce a weak, axisymmetric field with a large dipole offset. Further, the magnetic field, with remanent crustal magnetism indicating an ancient field in addition to the modern field, places limits on the rate of cooling over the planet's history.

Although explaining Mercury's weak magnetic field has long been a challenge (e.g., Heimpel *et al.*, 2005; Stanley *et al.*, 2005; Christensen, 2006), it is the combination of the weakness of the field and its axial alignment and asymmetry about the equator that makes understanding the dynamo mechanism even more intriguing. A common thread in many models of Mercury's magnetic field is the presence of a layer stable against convection (e.g., Christensen, 2006; Vilim *et al.*, 2010; Tian *et al.*, 2015). If such a layer is present, most likely at the top of the fluid core, then the heat flux out of the core may be less than what can be conducted along the adiabatic temperature gradient. In addition, compositional stratification may also be present. As discussed above, it is quite likely that there is a thermal component to the stability of such a layer, as thermal history calculations generally predict a subadiabatic heat flux at present. Furthermore, many potential core alloy compositions favor top-down crystallization regimes that lead to compositionally stratified layers. Thus, it seems likely that Mercury's core contains a stable layer that plays a role in the strength and geometry of the planet's magnetic field.

Yet the presence of a stable layer alone appears insufficient for explaining the strength and geometry of Mercury's magnetic field. To that end, recent models have included additional heterogeneity capable of further influencing magnetic field character (e.g., Figure 19.3). In particular, both Cao *et al*. (2014) and Tian *et al*. (2015) imposed laterally variable heat flux conditions at the CMB. Cao *et al.* (2014) utilized a heat flux pattern symmetric about the equator similar to the latitudinal variation in surface temperature consistent with Mercury's small axial tilt. Should the mantle be in a conductive, rather than convective state, then surface temperature variations at the surface may also have a signature at the CMB if enough time has passed since the end of the convective motions. Cao *et al*. (2014) investigated models with the highest and lowest heat flow at the equator, and they found that models with higher heat flow near the equator were better able to stabilize fields with geometries similar to those observed by MESSENGER. However, the mechanism for inducing larger heat fluxes along the equator, rather than at the poles, is unclear. Diffusion of surface temperatures to the CMB would result in relatively lower mantle temperatures near the poles, and therefore larger temperature differences and heat fluxes across the CMB there, rather than at the equator. As demonstrated in Figure 19.8, the limited thickness of the mantle seems to preclude long-wavelength convective patterns, so a degree-2 style of mantle convection appears unlikely as well. Therefore, some other mechanism for inducing a symmetric equator-to-pole variation in heat flux appears necessary for this mode of dynamo generation to operate.

Alternatively, Tian *et al*. (2015) imposed an asymmetric heat flux boundary condition along the CMB, with a higher heat flux out of the core near the north pole of Mercury (Figure 19.3). Those authors postulated that the higher heat flux there is a remnant of the magmatism that produced the NSP. As discussed in the previous section, there is a notable spatial dichotomy in the distribution of the youngest smooth plains on Mercury, with the largest expanses in the northern hemisphere (e.g., Ostrach *et al.*, 2015; Chapters 6, 11). However, as those volcanic deposits were emplaced 3.7–3.8 Ga, the thermal conditions that generated them are likely no longer present. Furthermore, smaller though still extensive ($>10^5$ km^2 area) (Byrne *et al.*, 2016) smooth plains units, the youngest effusive volcanic deposits on Mercury, are relatively well distributed between the northern and southern hemispheres, exclusive of the NSP and the plains associated with Caloris. Thus, it is worth considering whether the mechanisms for the large volcanic deposits and the smaller, more evenly distributed smooth plains deposits are the same (including whether some of the smaller deposits are even volcanic). Whereas the relatively larger concentrations of K at high northern latitudes on Mercury (Peplowski *et al.*, 2012; Chapter 7) might argue for a mantle source more enriched in heat-producing elements, such enhanced heat production would in fact lead to smaller temperature contrasts and a lower heat flux across the CMB. Interestingly, the K enhancement at high northern latitudes does not respect the morphological boundaries of the northern plains, nor are the lavas in Caloris so enriched. However, if the generation of the NSP substantially depleted the mantle at high northern latitudes of heat-producing elements compared with the rest of the planet, then core heat fluxes might be somewhat higher there due to the cooler mantle temperatures. The relatively limited amount of lateral mixing of the mantle expected under low-Rayleigh-number convection, coupled with the small scale of convection, could act to preserve such heterogeneity.

It is worth noting that MESSENGER gravity and topography data indicate that the domical rise within the northern volcanic plains is substantively compensated within ~100 km of the CMB (James *et al*., 2015). James *et al*. (2015) investigated a variety of mechanisms for the source of this compensation, including relief along a compositional interface (e.g., between the silicate mantle and a possible solid FeS layer at the top of the core) as well as other density variations. Variations in the thickness of an FeS layer would also result in changes in the thermal conductivity profile above the liquid core, leading to lateral differences in heat flux. A variety of compositions or viscosities at that depth may also induce additional thermal heterogeneity, though the impact of such variations relative to the remainder of the planet remains to be investigated.

It is clear that heterogeneity within Mercury's mantle may influence the mechanisms by which the planet's magnetic field is generated, though more work – and the need for further observations – remains. Indeed, any geochemical and petrologic heterogeneity (Chapters 2, 7) inherited from Mercury's earliest history may have substantially influenced the planet's history yet, as less is known about the geochemical and geophysical character of the entire southern hemisphere than the north, we have much more to learn about the distribution of any heterogeneous properties of the interior.

Mercury's internal structure and chemical make-up strongly influence the manner by which the planet's core, and therefore its magnetic field, has evolved. The discovery of Mercury's remanent crustal magnetism (Johnson *et al.*, 2015; Chapter 5) in crust that was last emplaced before ~3.7 Ga raises the question of how a planet cooling as modestly as suggested by its record of global contraction could have hosted both a relatively protracted period of early magnetic field generation and a modern field. A purely thermally generated dynamo that spans both time periods is unlikely, as the thermal history models indicate that core heat flux drops below the critical value for convection early in the planet's history and remains so. Indeed, early-onset thermal dynamos would tend to be short-lived, as evidenced by Figure 19.7 and previous modeling efforts (Hauck *et al*., 2004; Grott *et al*., 2011; Tosi *et al*., 2013). Although much shorter than the upper bound of ~800 Myr implied by the surface age of the crust in areas of remanent magnetism, such shorter-duration dynamos are potentially consistent with observations, as the column of crust hosting the remanence may pre-date the surface age. Models with longer-lived supercritical core heat fluxes are also possible. Under that scenario, the simplest explanation for the modern magnetic field is that it restarted comparatively recently as a result of the onset of core crystallization and perhaps even inner core growth. Alternatively, core crystallization could have been operating throughout the past 3.7 Gyr in order to account for both the ancient and modern fields. This mechanism is possible, yet would likely result in solidification of a substantial fraction of the core and greater contraction of the planet than has been documented so far. A large inner core does not appear to be compatible with the planet's internal structure (Hauck *et al.*, 2013; Dumberry and Rivoldini, 2015; Peale *et al.*, 2016) nor with magnetic field generation, as compositional gradients imposed by top-down crystallization, coupled with a large inner core, may serve to stabilize the entire core against convection (Dumberry and Rivoldini, 2015; Rückriemen *et al.*, 2015). Thus, a full understanding of the operation and evolution of Mercury's magnetic field depends on characterizing the age distribution of remanent crustal magnetism and understanding how core evolution, including the effects of core chemistry, was coupled to mantle convection and cooling through time.

19.7 Conclusions

MESSENGER has been instrumental in unveiling key elements of the global evolution of Mercury. From firmly establishing the occurrence of volcanism and its distribution in space and time, to substantively resolving the long-standing paradox between predicted and observed values for Mercury's global contraction and cooling, MESSENGER has brought new insight to fundamental questions about the planet that stood for nearly four decades. In turn, and as with all new missions of discovery, MESSENGER has raised new questions about how Mercury has operated over its history. With Mercury's remarkably thin mantle, which is incapable of significantly homogenizing its chemical character by mantle convection, it is clear that chemical heterogeneity has played an important role in the planet's history. The weak, axially aligned, and northward offset geometry of the internally generated magnetic field may be a distinct manifestation of internal heterogeneity. However, it is the discovery of Mercury's ancient magnetic field, recorded in the crustal rocks, that may hold some of the deepest clues to the planet's internal evolution.

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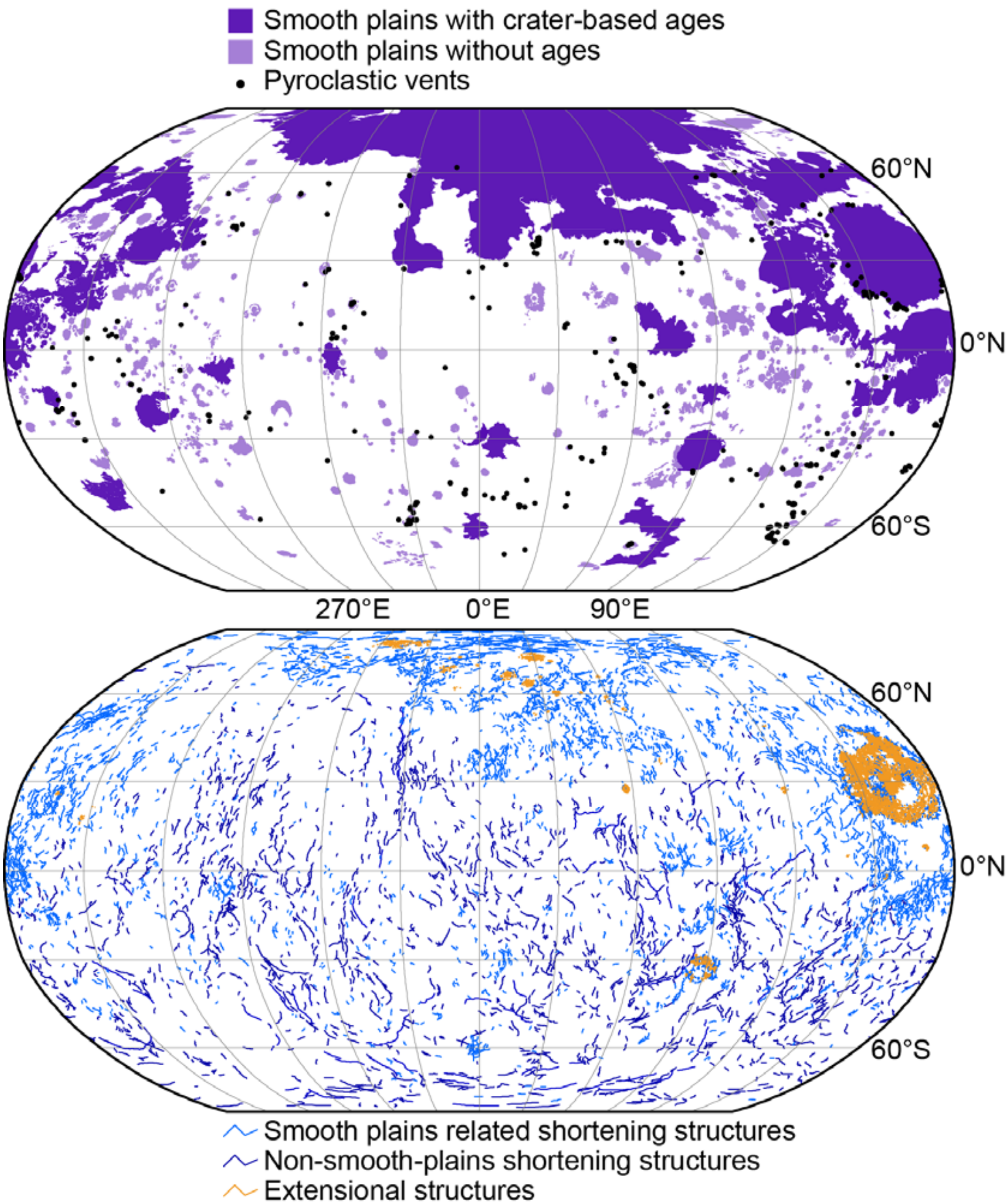

Figure 19.1. Overview of major geological features on Mercury. Top: Smooth plains are in purple; the darker units have estimated ages whereas the lighter shade units are too small for reliable crater-based ages. Mapped units are from Denevi *et al.* (2013) and Byrne *et al.* (2016). Locations of pyroclastic vents are from the compilation of Thomas *et al*. (2014). Bottom:

Compilation of tectonic structures, with shortening structures outlined in shades of blue and extensional structures in orange. Structures in the light blue are associated with smooth plains units as outlined in the top map in purple. The shortening structures are from Byrne *et al*. (2014), and the extensional structures are compiled from Klimczak *et al*. (2012), Ferrari *et al*. (2014), and Chapter 10.

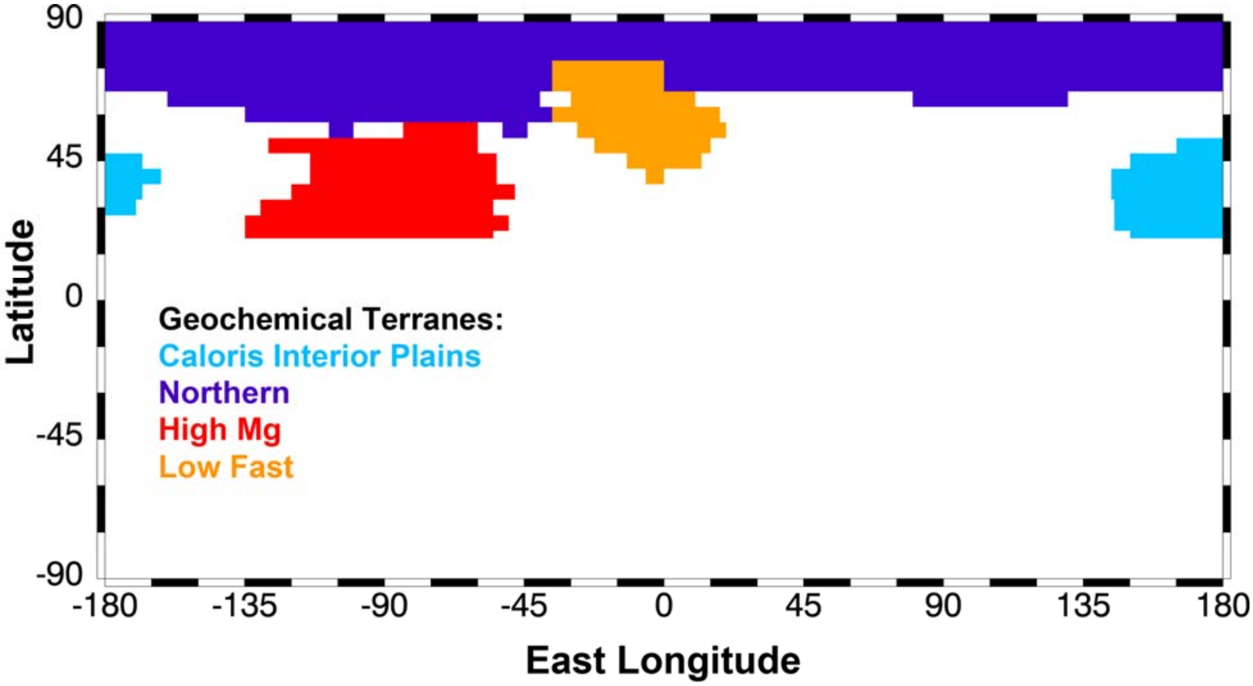

Figure 19.2. Overview of geochemically distinct terranes on Mercury (Chapters 2, 7). Figure from Patrick Peplowski.

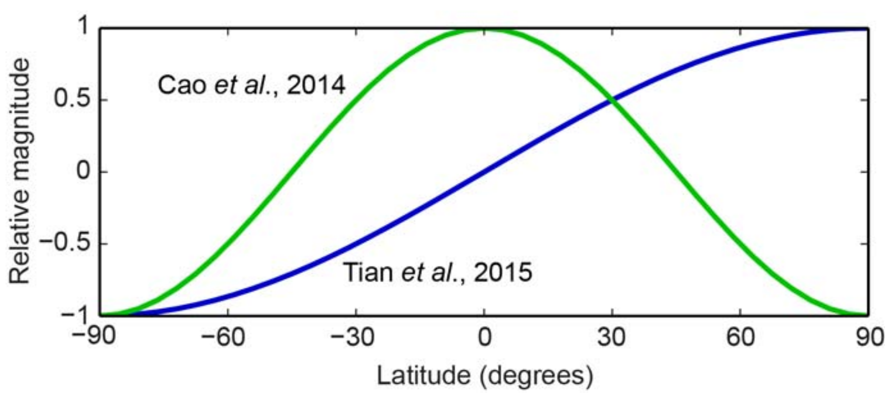

Figure 19.3. Relative variations in the imposed heat flux along the core–mantle boundary in MESSENGER-era dynamo models. The work of Cao *et al*. (2014) invoked a core heat flux that is higher at, and symmetric about, the equator, whereas Tian *et al*. (2015) assumed a core heat flux that is greater in the northern hemisphere than in the southern hemisphere.

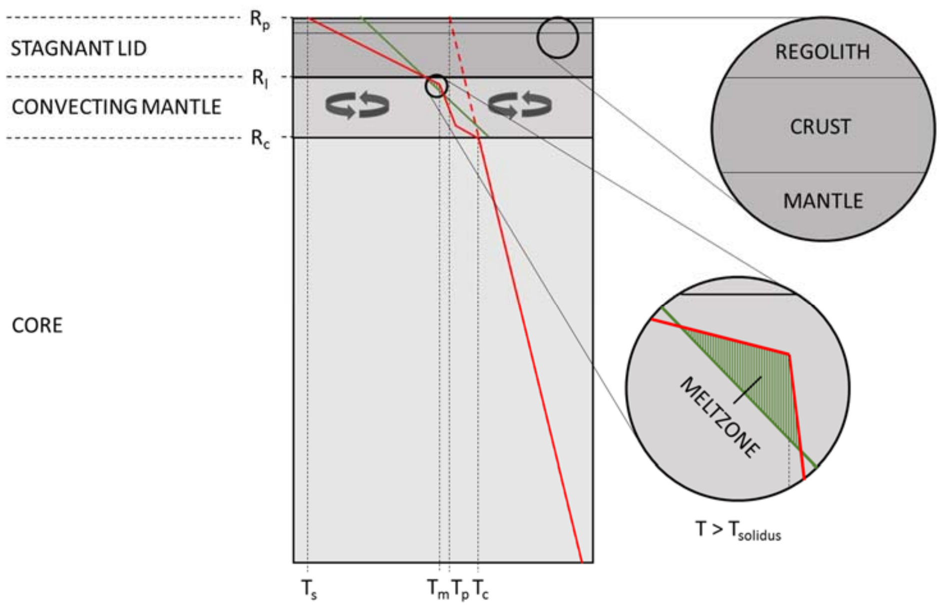

Figure 19.4 Schematic diagram of the reservoirs considered in parameterized thermal evolution models, including the stagnant lid, the convecting mantle, and the core. The planetary radius R_p, stagnant lid radius R_l, and core radius R_c are indicated. Temperatures shown are the surface temperature T_s, the upper mantle temperature T_m, the mantle potential temperature T_p, and the core temperature T_c, with temperatures increasing to the right. The mantle solidus is indicated schematically by the green line, and the melt zone in which the local temperature exceeds the solidus is indicated by the filled area in the lower inset.

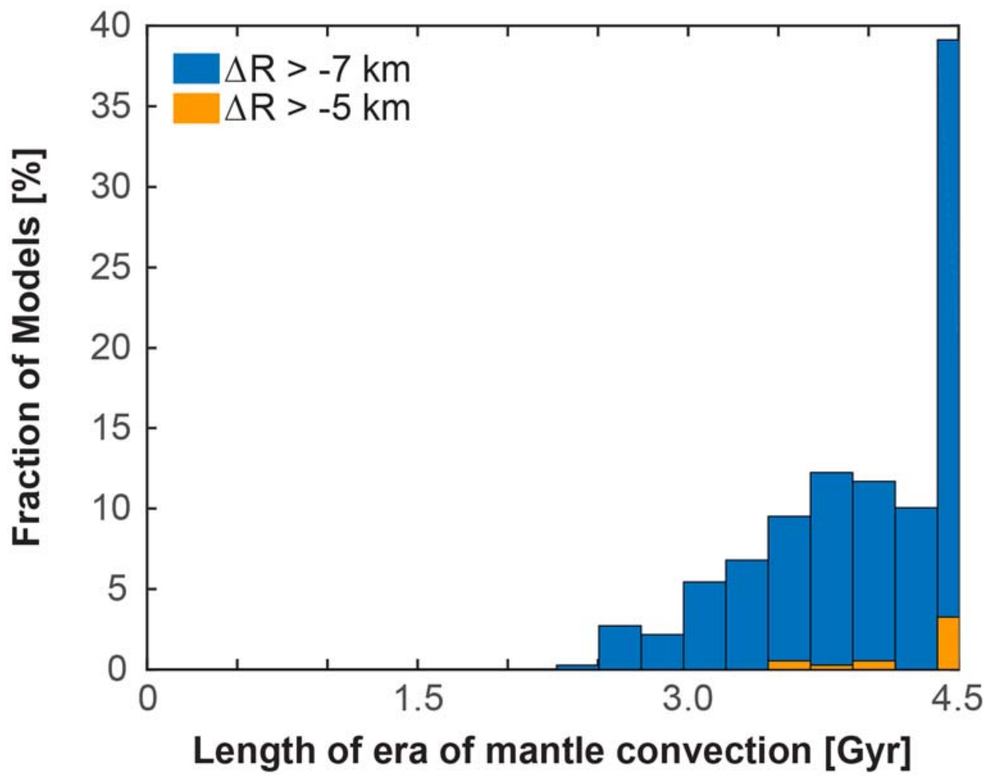

Figure 19.5. Results of Monte Carlo simulations of Mercury's thermal evolution for the duration of mantle convection. A total of 351 (blue) out of 2000 models from the simulations are consistent with the constraints posed by Mercury's magmatic evolution, global contraction, and magnetic field generation. The histogram shows the fraction of models in which mantle convection stopped at a given time. About 40% of the successful models convect to the present. Models shown in blue predict a reduction in planetary radius of between 5 to 7 km. This result should be compared with the models in orange, in which global contraction less than 5 km occurs but which otherwise satisfy the constraints, indicating the sensitivity of the inference on the longevity of mantle convection to the total observed radial contraction. Because of the uncertainty in core composition (section 19.4.4), contraction from inner core growth is neglected in these calculations. Note that the convention for global contraction here is a negative change in radius.

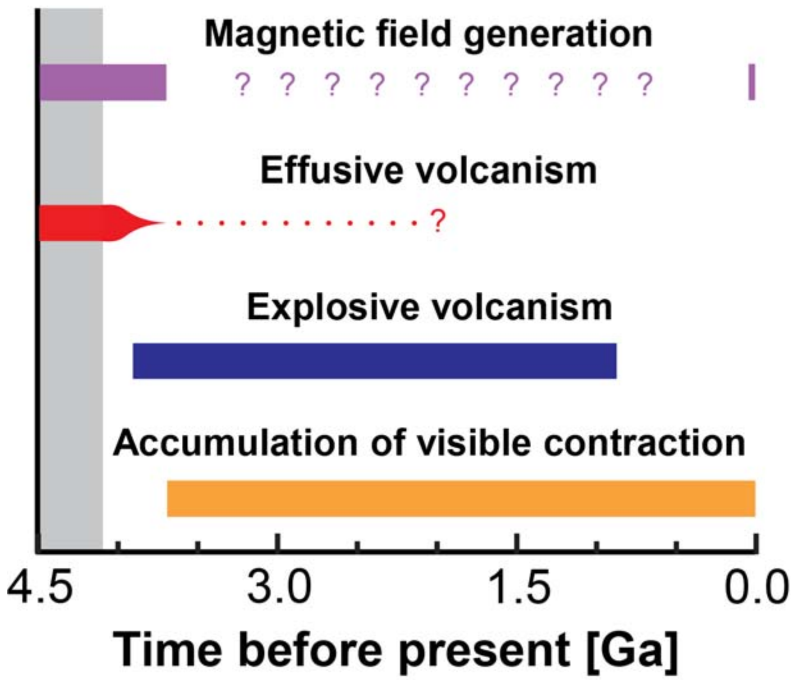

Figure 19.6. Schematic timeline of major processes in Mercury's evolution. Evidence of the planet's history during the first ~500 Myr has been erased by effusive volcanism and impact bombardment, as indicated by the gray shading.

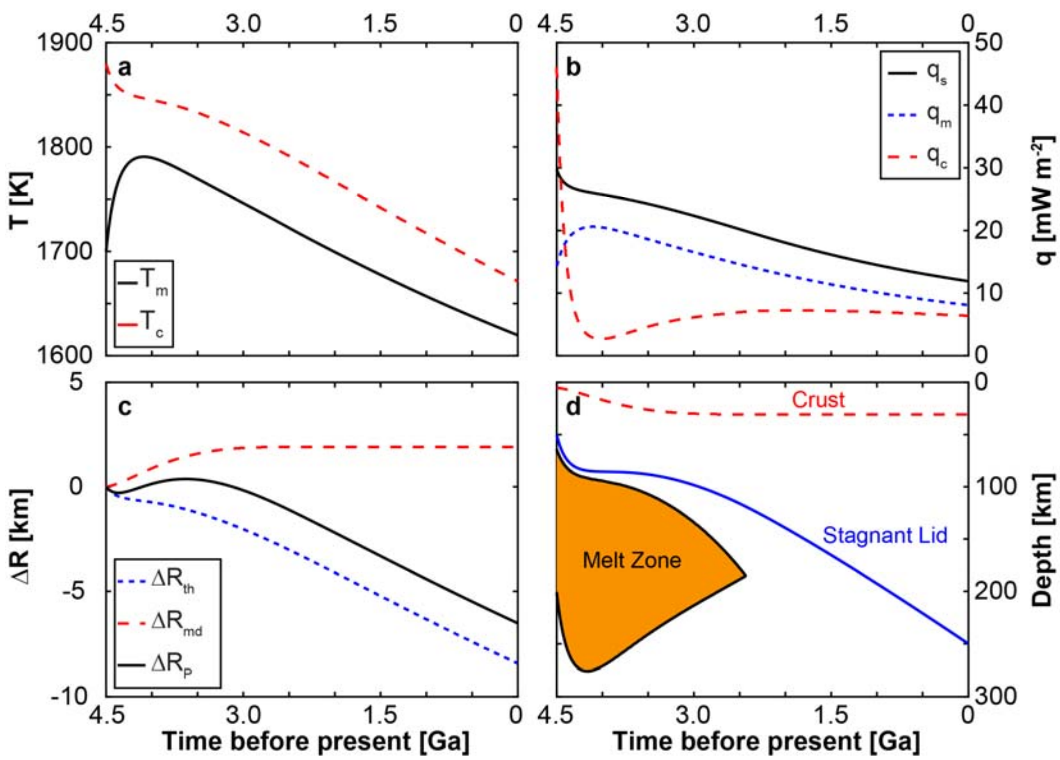

Figure 19.7. Representative thermo-chemical evolution model for Mercury with parameters as discussed in the text. (a) Evolution of mantle temperature T_m and CMB temperature T_c; (b) evolution of the surface heat flux q_s, mantle heat flux q_m, and core heat flux q_c; (c) evolution of the planetary radius change from thermal expansion and contraction of the mantle and core R_{th}, from mantle differentiation R_{md}, and of the sum of the two contributions R_P; (d) evolution of the thickness of the secondary crust, of the stagnant lid, and of the region in which partial melting occurs.

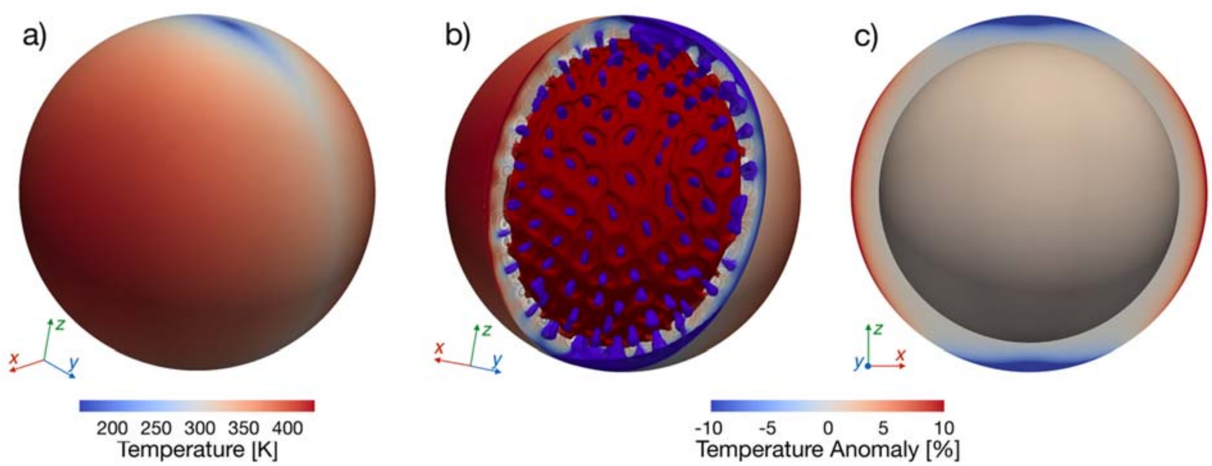

Figure 19.8. (a) Distribution of Mercury's average near-surface temperature according to the model of Vasavada *et al.* (1999). Hot equatorial poles are located at 0° and 180° E longitude, whereas cold poles are located at ±90° E. (b) Interior temperature anomalies after 1 Gyr of evolution when the mantle was still convecting. The color scale refers to the two mantle slices passing through the 0° and 90° meridional planes (the x–z and y–z planes, respectively), on top of which streamlines are plotted. Blue isovolumes mark the locations of downwelling that is 4 to 5% colder than average; red isovolumes refer to upwelling that is 1 to 2% hotter than average. (c) Interior temperature anomalies at present after the mantle transitioned to a conductive state, shown on the 0° meridional plane (x–z). Figure courtesy of Nicola Tosi.