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Question 3: Origin of Earth and Inner Solar System Bodies

How and when did the terrestrial planets, their moons, and the asteroids accrete, and what processes determined their initial properties? To what extent were outer Solar System materials incorporated?

Our inner Solar System has the four terrestrial planets, Mercury, Venus, Earth, and Mars, along with Earth's large Moon, two small moons of Mars, dwarf planet Ceres, numerous 100-km class asteroids, and a multitude of small bodies that populate the asteroid belt and the inner planet region. To what extent does this structure reflect a deterministic outcome of general Solar System formation processes, and how much of it is instead a consequence of stochastic events that would change in unpredictable ways if the system's formation could be rerun? This fundamental question motivates scientific investigations of the inner Solar System, with implications not only for understanding our own planet's origin but also for the origin of planetary systems and Earth-like planets beyond our own.

For example, general aspects of solid planet formation—including collisional accumulation from initially small planetesimals, temperatures too high for ice condensation in inner disk regions, and minimum orbital separations between large planets needed for stability—suggest that rocky inner planets similar to those formed in our Solar System might be a common outcome of terrestrial planet accretion. In contrast, processes such as giant planet migration and sporadic giant impacts may depend sensitively on conditions that vary greatly from system to system. The interplay between predictable and randomly determined events is an important theme in Solar System origin. Study of our inner Solar System worlds provides a uniquely valuable means to address this cross-cutting issue, thanks to their accessibility and the powerful constraints provided by the combination of remote observations and analyses of physical samples. Is our inner Solar System—and our Earth—a typical outcome of planetary system formation, or is it an outlier compared to most systems in the universe?

Q3.1 HOW AND WHEN DID ASTEROIDS AND INNER SOLAR SYSTEM PROTOPLANETS FORM?

The classic model for the formation of asteroids and inner Solar System protoplanets involves a transition from a proto-planetary gas nebula characterized by dust and gas to one populated by hundreds of 100 km sized bodies. Meteorites analyzed on Earth and asteroids studied in space preserve some of the compositional, mineralogical, and isotopic characteristics of these early formed bodies. The study of these remnants of Solar System formation are a critical component for constraining both the physical and chemical characteristics of the terrestrial planets, as well as for understanding the processes that occurred in the protoplanetary gas nebula. However, many fundamental questions remain unanswered. We do not understand the origin of the materials these planetesimals derive from, or even if the current population of meteorites and asteroids is representative of the population that comprise inner solar system terrestrial bodies. Similarly, although we know that these small bodies undergo varying degrees of heating and associated aqueous alteration and differentiation, we are unsure how this translates into compositional variations observed in larger bodies. In fact, there are even gaps in our knowledge of how accretion works in the context of an ever evolving protoplanetary gas nebula. Addressing these questions is fundamental to a better understanding of how the Solar System evolved into its present state.

Q3.1a What were the feedstocks to the early inner Solar System and did their compositions change with time?

The compositions of planets are set by an array of processes, perhaps the most fundamental being compositional variations in the initial materials. The solar nebula was a dynamic environment, with distinct compositional reservoirs developed early in Solar System history. Many properties of these reservoirs may reflect differences in distance from the Sun (Question 1), while others would have been affected by early disk, planetesimal, and/or protoplanet growth conditions. Isotopic, chemical, and mineralogical heterogeneities in the building blocks of the inner Solar System helped produce the diversity of inner Solar System bodies seen today, which show strong variations in composition.

The innermost planet, Mercury, has a highly reduced surface, with low iron contents and unexpectedly high sulfur contents, consistent with a planet formed from highly reduced materials. Information about the composition of Venus is extremely limited, a fundamental gap in our understanding of the compositional variations among the terrestrial planets. Studies indicate the Earth may have formed from more reduced materials in its early accretion stages, followed by the agglomeration of more oxidized materials. Mars, the outermost terrestrial planet, has a more oxidized surface and (by some arguments) interior. This could reflect a radial gradient in oxidation fugacity among planetesimals in the early Solar System, which in turn would imply limited dynamical mixing among early planetesimal reservoirs, such that the terrestrial planets mainly formed from local materials in the protoplanetary disk.

In addition, two reservoirs of planetary building blocks have been identified through the analysis of primitive meteorites. The first is associated with carbonaceous chondrites (CC) and the second with non-carbonaceous chondrites (NC). The CC reservoir is characterized by more volatile-rich compositions and a higher proportion of isotopes produced by explosive nucleosynthesis, i.e., supernovae, prior to Solar System formation, whereas the NC reservoir is less volatile element-rich and has a greater proportion of isotopes produced by fusion in the cores of stars prior to Solar System formation. The differences in the nucleosynthetic isotope signatures of these reservoirs, combined with the inference that volatile element abundances reflect condensation location within the protoplanetary disk, indicate the disk was radially stratified at the time of accretion. The Earth appears to represent a compositional endmember, suggesting that a reservoir, representing the Solar System inside Earth's orbit, may not yet have been identified or sampled (Dauphas 2017; Mezger et al., 2020). Samples from Mercury and/or Venus would be invaluable in helping to reconstruct the conditions of inner Solar System accretion.

Early alteration processes, such as chemical and physical modification of gas and dust by the young Sun, acting on nebular materials may have been important. Depending on the timing of the initiation of hydrogen fusion in the Sun, this process could have bathed the surrounding gas and dust in light, including ultraviolet radiation, causing chemical changes. As material aggregated into pebbles and planetesimals (see Question 1), this energy, in combination with energy released by radiogenic decay of short-lived isotopes and gravitational energy, would have melted some solids and liberated volatile elements that could enable aqueous alteration of materials, fundamentally altering building block compositions. Likewise, melting and internal differentiation of planetesimals could also have led to significant differences in the chemical and physical characteristics of subsequently accreted planetary bodies. For example, a planet formed from CC material might be expected to be volatile rich and oxidized (Q3.6). However, if the CC material first underwent melting and differentiation in a planetesimal, a resulting planet could be more reduced and have lower abundances of volatile elements and compounds due to volatile loss during the planetesimal phase. Constraining the effects of the alteration and differentiation process on the composition of planetary bodies relies on understanding leftover materials from within the Solar System, such as primitive chondrites and small planetesimal bodies. The range of evolutionary processes incorporating these ideas are discussed further in section Q3.1c.

Some of the best constraints on the building blocks of the terrestrial planets come from meteorite samples and asteroid studies. Asteroids display considerable diversity in spectral properties (e.g., color, albedo), size, and geophysical characteristics, and may include materials that originated across both the inner and outer Solar System (see Q3.1c). While there are a multitude of asteroids, their total mass is small, 0.04% of Earth's mass, and the extent to which their compositions are fully representative of materials that formed the terrestrial planets remains unclear. More reduced inner Solar System materials may have been lost to accretion, although perhaps some were scattered into the asteroid belt during planet formation or were captured as Mars Trojans. Among reduced meteorite samples, there are hints of these missing building blocks and a suggestion that there was less mixing between different inner Solar System planetary zones than previously thought.

Samples of the Earth provide a foundation for all compositional studies, and we are fortunate to have samples from the Moon, specific asteroids (including returned samples from (25143) Itokawa, and soon (101955) Bennu and (162173) Ryugu), meteorites from Mars, and numerous meteorites from other meteorite-parent-body asteroids that enable continual advances in compositional measurements.. Samples from Mercury and Venus have hitherto not been obtained or recognized if they exist within our meteorite collection. Refined in situ geochemical characterization of surface materials at Mercury and Venus would transform our knowledge of these bodies, especially in the absence of samples. Although we have a multitude of asteroidal meteorites, extraterrestrial materials delivered naturally to the Earth are strongly biased: for example, icy materials do not survive during transit through Earth's atmosphere. There potentially remain unsampled asteroidal materials that could be investigated by compositional measurements or sample return from rare or unexplored asteroid spectral types.

Q3.1b What were the mechanisms of accretion from planetesimals to larger bodies?

The standard story of planet formation, told for decades, is that planets formed from the solar nebula cloud of dust and gas in a multi-part process termed accretion (Question 1). The first solids to condense from nebular gasses were composed predominantly of refractory elements and include Calcium-Aluminum-rich inclusions (CAIs) and more silicate-rich chondrules. These early-formed solids subsequently accreted into small, perhaps 100-km scale planetesimals. In the inner Solar System, planetesimals collisionally accreted into Moon-to-Mars sized planetary embryos within a few million years. Planetary embryos initially on nearly circular, co-planar orbits began to gravitationally perturb one another once the local mass in planetesimals and embryos became comparable and/or the gaseous nebula dispersed. Orbital eccentricities increased, leading to a phase of orbit crossing and giant collisions over tens to a hundred million years, reducing the number of protoplanets through mergers until only a few planets remained on stable orbits.

The standard model is popular because in numerical simulations it yields a modestly successful set of terrestrial planet analogs. When examined in detail, however, the model planets frequently have the wrong masses, orbits, and compositions. In particular, the mass of Mars analogs often rivals those of Earth and Venus, while real Mars has only about 10% the mass of the Earth. The "small Mars" problem is a driver for considerable innovation in current planet formation models. The inability of the numerical simulations to account for many fundamental characteristics of the Solar System suggests that we do not fully understand the processes by which dust and pebbles accrete to form the current set of planets and small bodies. It probably also indicates that we do not understand the dynamic interplay between nebular gas, dust, nascent planetesimals, and the Sun.

Q3.1c What evolutionary processes led to the initial diversity of asteroids and on what timescales?

According to planet formation and giant planet migration models, the main asteroid belt is a collection zone for planetesimals formed from across the Solar System. While some asteroids may have originated in the main belt region, the majority likely came from the terrestrial planet zone (predominately the Mars region), the giant planet zone, and the primordial Kuiper belt. This wide variety of source regions means the asteroids and meteorites potentially reflect a broad range of planetesimal compositions.

Meteorite studies have shown that the decay of short-lived radioisotopes resulted in heating, melting, and in some cases differentiation, of early formed planetesimals within the first few million years of Solar System history. Chronologic investigations further suggest that differentiation on parent bodies was contemporaneous, and in some cases earlier, than condensation of some chondrules thought to be one of the earliest solids to form in the Solar System. Thus, at least some of the building blocks of the terrestrial planets and giant planet cores are likely to have experienced melting and differentiation prior to their incorporation into these bodies. However, meteorites display the gamut of heating histories, manifested by varying degrees of alteration and differentiation. It is not clear if there is a continuum of heating and melting related to planetesimal formation ages or sizes, or if other events such as early collisions set different planetesimals on alternate evolutionary paths.

Collisions have shattered and scrambled asteroids over billions of years, creating mixed-up bodies with originally deep interior materials now on their surfaces or within ejected fragments. An example might be the iron-rich asteroid (16) Psyche, whose origin will be explored by the Psyche mission. Asteroid impacts can also mix projectile and target so that some worlds appear to be derived from materials sourced from different locations in the Solar System.

Spectral observations of asteroids have identified a wide range of classifications, a few of which have been linked to meteorite types by spacecraft exploration. For example, Hayabusa's sample return from (25143) Itokawa has demonstrated a link to ordinary chondrite meteorites, and Dawn's exploration of (4) Vesta has confirmed that (4) Vesta is the parent-body of the howardite, eucrite, and diogenite (HED) meteorites. Other associations between meteorite types and asteroid classes remain to be tested or discovered. For example, the spectral signature of a partially differentiated asteroid is completely unknown. Meteorite samples suggest that many asteroid bodies experienced complete differentiation, yet materials that represent mantles of differentiated asteroids are extremely rare, in both our meteorite collections and in asteroid observations. This may be due to processes preferentially breaking these materials up or changing their spectral signatures through space weathering (a collection of processes that alter the surfaces of material when exposed to the space radiation environment, as described in Q5.5). Linking the diversity observed in meteorite samples to the large-scale diversity observed in the asteroid population is fundamental to advancing our knowledge of both types of objects.

Even primitive chondritic meteorites—thought to best reflect the initial, unaltered composition of the protoplanetary disk—show evidence of parent-body processes that have affected their properties and mineralogy. Hydrothermal alteration, thermal metamorphism, shock heating, and compaction occurred early in the Solar System, prior to complete accretion of the terrestrial planets. All of these processes played important roles in determining the physical characteristics of chondritic materials and contributed to the variability of the building blocks of terrestrial bodies. As one example, although there is evidence that aqueous alteration can affect the abundances and chirality of amino acids in primitive materials that were potentially delivered to Earth, there are open questions about the nature and extent of aqueous alteration on these bodies (Chapter 5), such as the peak temperatures, lifetimes of liquid water, and the role of gas phases. Understanding the range and extent of these evolutionary processes is important to interpreting the nature of the asteroid population and for constraining the building blocks that formed the inner planets.

Strategic Research to address Q3.1:

- **Determine the compositional diversity of the terrestrial planets and inner Solar System feedstocks** by obtaining mineralogical, geochemical, and isotopic data from the surfaces and atmospheres of Mercury, Venus, Moon, and the less explored regions of the Moon and Mars, as well as the currently unsampled small body population.
- **Determine the diversity of compositions and nature of remnant planetesimals residing in the inner Solar System and establish links between the small body taxonomy and meteorite types** through Earth-based and spacecraft-based remote sensing, in situ measurements, and laboratory analyses of meteorites and returned samples.
- **Evaluate the nature of early projectiles that struck the terrestrial planets and the Moon** by analyzing regolith samples likely to contain remnant clasts from early bombardment impactors and by obtaining isotopic traces of projectile materials from lunar craters and basins.
- **Determine temporal changes to inner Solar system feedstock compositions** by coupling geochronological measurements with other geochemical and isotopic measurements for refractory- and volatile-rich materials from a wide range of parent bodies.
- **Determine what secondary processes have led to the diversity of asteroids and planetary feedstocks** by conducting geochemical, petrological, and geophysical investigations of meteorites, asteroids, and samples returned from asteroids.
- **Determine the mechanisms of planetary accretion** by developing and evaluating physical models coupled with observational constraints on compositions and distributions of material and planetesimal structures comprising the Solar System.

Q3.2 DID GIANT PLANET FORMATION AND MIGRATION SHAPE THE FORMATION OF THE INNER SOLAR SYSTEM?

Historically, it was thought that the giant planets and small bodies formed near their current locations, but it has become increasingly clear that the structure of the Solar System changed as the giant planets grew and their orbits migrated. This migration could have been driven by interactions between the giant planets and the gas nebula, or by later interactions with leftover planetesimals after the gas nebula dissipated, or both (see Question 2). Depending on when and how it occurred, giant planet migration could have reshaped the terrestrial planet region as the inner Solar System was forming.

Q3.2a How would early, nebula-driven migration have affected the inner solar system?

The earliest phase of giant planet migration may have occurred within the protoplanetary gas nebula. In the best developed model of this behavior, the so-called “Grand Tack” model, gas disk interactions led early Jupiter to migrate *inward* to ~1.5 AU from the Sun, at which point Jupiter and Saturn became trapped in a mutual 2:3 resonance and began to move *outward* (Walsh et al. 2011). Outward migration continued until the gas disk dispersed, presumably when Jupiter reached its current distance of ~5.5 AU. The net effect of the migration was to deplete the Mars zone of material, providing an explanation for the small mass of Mars compared to Earth and Venus (Hansen 2009).

A key feature of the Grant Tack is that numerous planetesimals are scattered into the inner solar system from the giant planet zone. Some are captured into the central and outer main asteroid belt, the observed location of the carbonaceous chondrite asteroids (CC bodies). This would link CC asteroids to giant planet zone planetesimals. Other outer Solar System planetesimals may have struck the growing terrestrial planets, providing a source of water-rich material.

There are other, non-Grand Tack scenarios to explain Mars’s small mass (e.g., Q3.2b), and how CC bodies may have been delivered to the inner Solar System. A key question is whether any CC bodies were formed originally in the main belt. NC and CC meteorites have major differences in titanium, chromium, oxygen, molybdenum, and tungsten isotopic compositions, and this has been interpreted to mean that NC and CC planetesimals had to form in distinct chemical reservoirs. One way to produce and maintain such reservoirs is if Jupiter formed early and acted as a barrier against inward migration of material (Kruijer et al., 2017). In this case, NC and CC bodies would represent inner and outer solar system planetesimals, respectively, implying that the only NC asteroids were originally formed in the main belt. Future work should focus on testing these scenarios by seeking evidence that Jupiter and Saturn migrated across the primordial main asteroid belt, by defining the mass of the asteroid belt prior to giant planet migration, and the fate of bodies potentially lost from the asteroid belt during this migration. Tests of these scenarios will require meteorite studies, ground-based and space-based observations, data returned by small body missions, and modeling.

Q3.2b How was the inner solar system affected by giant planet migration after the gas disk dispersed?

The orbital properties of Pluto and other Kuiper Belt objects provide compelling evidence that early Neptune migrated substantially outward (see Question 2). Decades of theoretical models have shown that after the protoplanetary gas nebula dispersed, the orbits of Saturn, Uranus and Neptune would have indeed migrated outward due to gravitational interactions with a remnant outer planetesimal disk, while such interactions would have caused Jupiter’s orbit to contract. During such migration the giant planets often undergo an orbital instability, such as in the widely explored “Nice” model (Gomes et al 2005; Figure 3.1).

A giant planet instability would have affected the terrestrial planet region, depending on when it occurred. The timing of the instability is uncertain, because it is controlled by the unknown size of the gap between Neptune’s initial orbit and the primordial Kuiper belt’s inner edge. The smaller the gap, the earlier the instability. If the gap was <1.5 AU, Neptune's migration would take place shortly after dissipation of the protoplanetary gas nebula, within the first few to 10 million years of Solar System history. This migration would then have had a major influence on terrestrial planet accretion, and could provide an alternative to the Grand Tack for explaining Mars' small mass, with Mars zone planetesimals scattered away before Mars could fully form. Alternatively, a late instability after terrestrial accretion was complete—initially favored as a way to explain a clustering of lunar basin ages at about 4 billion years ago (the so-called "late heavy bombardment" or terminal lunar cataclysm)—would have likely destabilized the orbits of the fully-grown terrestrial planets. To date, only a small fraction of existing simulations find that the terrestrial planets could avoid this fate, thus favoring an earlier instability before the terrestrial planets reached their final configuration (see Bottke and Norman, 2017).

The number of giant planets gravitationally interacting with one another at the time of the instability is also unknown. It is possible the Solar System had additional gas giants early in its history that were subsequently ejected. Dynamical studies suggest systems with 5 or 6 giant planets (e.g., 3 or 4 Neptune-size bodies) have greater success in reproducing dynamical constraints across the Solar System than those that start with 4 giant planets. In the most successful models, a Neptune-sized body interacting with Jupiter causes it to migrate slightly inward via numerous tiny jumps before it is eventually ejected, and these Jupiter "jumping events" strongly affect the dynamical structure of the asteroid belt Vokrouhlický, et. al.. (2016)..

Giant planet encounters with one another may also allow a small fraction of comets ejected from the primordial Kuiper belt to be captured in stable orbits across the Solar System. This might explain why primitive, comet-like asteroids (D- and P-types) are found in the central and outer main belt as well as in the Hilda and Trojan populations associated with Jupiter, and would imply that the asteroid belt contains planetesimals from the furthest reaches of the Solar System. As the giant planets moved to their present-day orbits, dynamical resonances associated with them also moved to their current locations, perhaps depleting or eliminating portions of the primordial inner main belt while perhaps creating the innermost asteroid population , the Hungaria asteroids.

An overall goal is to find evidence for or against these dynamical set pieces through missions and/or meteorite analysis. We need to determine precisely how the signatures of post-nebula giant planet migration are recorded in small body populations and whether the nature of the asteroid belt can tell us how many giant planets existed prior to the giant planet instability. We want to determine whether dormant comet-like asteroids were implanted in the main belt, Hildas, and/or Trojans by the giant planet instability and whether the primordial asteroid belt interacted with an "extra Neptune".

<u>Strategic Research to address Q3.2:</u>

- **Determine whether Kuiper belt objects, Centaurs, comets, and P- and D-type asteroids within the main belt, Hilda, Trojan asteroid, and irregular satellite populations originated in the primordial Kuiper belt** by measuring their properties via remote observations, in situ studies, and/or sample return.
- **Determine whether C- and B-type asteroids within the main asteroid belt originated within the giant planet region** by assessing their volatile content using in situ methods or sample return, and identifying whether any provide evidence of parent body origin at low temperatures beyond Jupiter.
- **Determine the timing of the giant planet instability** through evidence of early comet bombardment of the asteroid belt (e.g., impact history of large asteroids that resisted disruption, identification of common shattering/disruption times for asteroids from meteorite shock degassing ages) and constraining the ages of the oldest lunar impact basins.
- **Investigate how planetesimals from the giant planet zone may have reached the inner solar system in the presence of the gaseous nebula** through dynamical models coupled to constraints provided by the orbits and sizes of C- and B-type asteroids within the main belt and the abundance and distribution of inner Solar System volatiles.
- **Investigate how giant planet migration after gas nebula dissipation affected the asteroid belt and terrestrial planet accretion** by observing comet-like asteroids (i.e., D- and P-type bodies) within the main asteroid belt and using their properties to constrain if and how migration led to

their capture into the main belt, and through dynamical and collisional models coupled to compositional constraints.

- **Constrain the orbit of Jupiter prior to its post-nebula giant planet migration phase, and thereby the nature of the giant planet instability** by identifying putative asteroids that were initially captured in resonances associated with Jupiter's orbit prior to migration.

Q3.3 HOW DID THE EARTH-MOON SYSTEM FORM?

Among our terrestrial planets, only Earth has a large satellite, and the Moon has affected our planet throughout its history. Lunar and terrestrial samples, together with spacecraft data, provide an extensive and powerful set of constraints on the origin of Earth-Moon, making this system unique in its ability to reveal conditions during the final stages of terrestrial planet accretion. The favored giant impact hypothesis proposes that the Moon formed as a result of a collision between the proto-Earth and another protoplanetary embryo, “Theia”, at the end of Earth’s main accretion (Hartmann and Davis, 1975). The impact left the Earth with its rapid early rotation rate, and produced a disk of iron-poor, highly heated debris, which later accreted into the Moon forming a primordial lunar magma ocean (LMO). Although this general model accounts for many characteristics of the Earth-Moon system, some characteristics have proved extremely challenging to explain, spurring development of a wide range of new impact models that are actively debated. Questions also remain regarding the timing of the giant impact.

Q3.3a What are physical constraints on Earth-Moon system origin?

Key constraints on how the Earth and Moon formed include their compositional relationships, the Moon’s age, and its initial thermal state. The extent of understanding of each of these constraints has evolved substantially in the past decade, yet many open questions remain.

Laboratory analyses continue to provide breakthroughs in understanding of how the compositions of lunar samples relate to those of terrestrial rocks. Much recent work focuses on isotopic compositions thought to best reflect the original feedstock of both bodies. Remarkably, the silicate Earth and Moon have nearly identical isotopic compositions for many elements (e.g., O, Ti, Cr, Si, W), implying that they formed from a common source that was isotopically distinct from nearly all meteorite classes. The tungsten isotopic similarity is the most challenging to explain, requiring either a compositional match between the proto-Earth and Theia that is very improbable given their separate core formation histories, or complete mixing of protolunar material with the Earth’s mantle, potentially within vaporized phases produced during or after the impact. Better understanding of how late accretion (Q3.5d) affected Earth-Moon tungsten isotopic compositions, as well as how lunar and terrestrial rocks compare in highly refractory elements (e.g., Ca) less likely to vaporize, is important to further progress (Zhang et al 2012). Debate continues as to whether the Earth and Moon have identical oxygen isotopic compositions, or whether small differences between the two indicate that the Moon still partially records the composition of Theia (e.g., Young et al. 2016; Cano et al. 2020).

The Moon’s volatile content may provide important constraints on the energetics of the giant impact and the conditions of the Moon’s formation. Lunar samples are generally dry and depleted in volatile elements compared with terrestrial rocks, which are strongly volatile element depleted relative to most other planetary materials that have been analyzed. However, analyses of a small number of lunar samples, notably volcanic glass beads that trapped mantle melt, now show that at least portions of the early Moon were comparably water-rich to the Earth’s mantle (e.g., Hauri et al. 2015). The stable isotope composition of moderately volatile elements is slightly heavier and significantly more variable in lunar rocks than in terrestrial rocks (e.g., Wang and Jacobsen 2016). Accounting for these observations is challenging and has led to a variety of models of evaporative loss or partial condensation in the pre-lunar disk or by geologic processing on the early Moon (e.g., Canup et al. 2015; Day et al., 2017; Kato & Moynier 2017; Wang et al. 2019; Nie & Dauphas 2019).

An overall question is whether the Moon’s composition is heterogeneous with depth, which would imply that lunar samples are not representative of the Moon’s bulk composition. Improved constraints on the Moon’s bulk composition, through additional samples originating in its deep interior (volcanic glasses or mantle exposed from the largest basin-forming impacts, particularly South Pole–Aitken basin) and/or a more thorough understanding of the Moon's geophysical structure, would be of great value.

Current estimates for the age of the Moon forming giant impact range from 4.52 to about 4.42 billion years ago, corresponding to about 50 to 150 million years after formation of the earliest Solar System solids (e.g., Connelly and Bizzarro, 2012; Kruijer and Kleine, 2017, Thiemens et al. 2019; Maurice et al., 2020). The large uncertainty stems from the fact that these estimates are based on assumption-laden models of isotopic evolution, which must be assessed in the context of multiple geologically related events such as accretion, differentiation, and lunar magma ocean solidification (see Q3.6 and Question 5).

A fundamental issue for origin models is whether the Moon was completely or only partially molten when it formed—a so-called magma ocean. Multiple geophysical constraints seem to imply an initial solid lunar interior beneath the lunar magma ocean, including the history of tectonics and strain most recently revealed by GRAIL data (Andrews-Hanna et al. 2013), and a seismic transition that may represent the magma ocean base (Khan et al. 2006). However, alternative explanations exist that could permit a fully molten Moon given current data limitations and uncertainties. Resolving this issue is important because a partially molten Moon is very difficult to reconcile with a giant impact origin, and thus would provide strong constraints on lunar origin models. Long temporal baseline geophysical and seismic data are needed to reveal the Moon's interior structure and its early thermal state. Geochemical analyses of additional and more diverse crustal samples, as well as improved constraints on the Moon's bulk composition, would also contribute significantly to this fundamental issue.

Q3.3b What was the nature of a Moon-forming giant impact(s) and its implications for the initial states of the Earth and Moon?

A canonical Moon-forming impact by a Mars-sized Theia can account for the masses of the Earth and Moon, and the angular momentum of the system (e.g., Canup 2004). However, it produces a proto-lunar disk that originates primarily from Theia rather than from the proto-Earth. Meteorites that originate from Mars, and nearly all those from the asteroid belt, have substantially different isotopic compositions than the Earth. If Theia were similarly non-Earth like, one would expect measurable differences between the Earth and Moon. Instead, the Earth and Moon are nearly isotopically indistinguishable for all non-volatile elements. A multitude of new concepts have been proposed to resolve this so-called "isotopic crisis" for the giant impact model. Notable examples include mixing and chemical equilibration of the silicate Earth with pre-lunar material (Pahlevan and Stevenson 2007; Lock et al. 2018); high angular momentum impacts (Cuk and Stewart 2012; Canup 2012; Lock et al. 2018); a "hit-and-run" impact (Reufer et al. 2012); formation of the Moon by multiple impacts (Rufu et al., 2017); and a Theia and proto-Earth that had similar isotopic compositions (Dauphas, 2017).

The relative merit of these different models is a matter of active debate (e.g., Canup et al. 2021), and progress will require improved understanding on multiple fronts. First, current thinking that Theia would have been isotopically distinct from the Earth is heavily influenced by the isotopic composition of martian meteorites, the only other inner planet from which we have samples. Knowledge of the isotopic compositions of Venus and/or Mercury—currently unknown—is needed to address this fundamental issue and reveal the primordial feedstock of our innermost planets. Whether lunar materials could have thoroughly mixed with the Earth's mantle prior to the Moon's assembly remains unclear, and needs to be assessed through modeling and observational tests.

Another central issue is whether the Earth-Moon system angular momentum was greatly modified after the Moon formed. Tidal interaction between the Earth and Moon conserves their total angular momentum, but certain gravitational interactions involving the Sun could have slowed the Earth's spin and reduced the system angular momentum by a factor of 2 or more soon after the Moon formed (e.g., Ćuk and Stewart 2012). If this occurred, it implies the Moon-forming giant impact was vastly more energetic than originally envisioned. Uncertainty persists because the spin-slowing mechanisms depend sensitively on the nature of early tidal dissipation in both bodies, in turn a function of their early thermal evolution. Additional constraints may arise from better understanding of the Moon's early orbital evolution and the related origin of its fossil figure (the remnant shape of the Moon, frozen in from this early epoch; Garrick-Bethell et al. 2014, Keane & Matsuyama 2014).

Overall, unraveling how the Earth-Moon formed will require development of improved observational tests to allow model predictions—e.g., for the Moon's composition and initial thermal state—to be compared with past and future data. Primary questions involve the physical processes and mixing associated with the giant impact and their effects on the proto-Earth, conditions in the pre-lunar disk produced by the impact (including whether it was ionized and affected by magnetic fields), and the nature of subsequent lunar accretion and early orbital evolution.

Strategic Research to address Q3.3:

- **Determine the internal structure of the Moon with sufficient resolution to constrain its bulk composition and initial thermal state** using geophysical measurements obtained from spacecraft and/or a seismic network and other in situ analyses.
- **Determine the Moon's interior composition** by sample return and/or in situ analysis of materials that reflect the Moon's endogenic composition at depth, e.g., glass beads, primitive basalts, and/or exposed lunar mantle.
- **Constrain the physical and chemical characteristics of Theia and the proto-Earth** through sample analysis or in-situ isotopic measurements of inner Solar System bodies (particularly Venus or Mercury), as well as improved models to explain the isotopic and geochemical constraints of the Earth and the Moon.
- **Determine the origin of the Moon's volatile element abundances** by completing additional stable isotopic analyses of volatile elements in lunar samples originating from the lunar interior as well as gamma-ray and neutron spectrometer measurements (e.g., K/Th ratios) by spacecraft.
- **Determine the timing of the Moon-forming giant impact and solidification of LMO** by isotopic analysis of lunar rocks from nearside and farside of the Moon and by refining theoretical models to estimate the timescale for the duration of LMO crystallization.
- **Seek evidence for post-giant impact equilibrium between Earth and Moon** by analyzing terrestrial and lunar samples for stable refractory element isotopic compositions.
- **Differentiate between giant impact concepts** by developing model predictions for observable properties of the Moon and comparing them with lunar compositional and geophysical data.

Q3.4 WHAT PROCESSES YIELDED MARS, VENUS, AND MERCURY AND THEIR VARIED INITIAL STATES?

The four terrestrial planets, Mercury, Venus, Earth and Mars, share many similarities. All are solid bodies with a central core composed dominantly of metallic iron surrounded by layers of silicate rocks. However, there are substantial differences among these bodies that, at least in part, appear to reflect substantial differences in their formation histories.

Q3.4a Why are Earth and Venus so different?

The contrast between the current Earth and Venus is striking. Earth is an oasis for life, with plentiful liquid water on its surface, a temperate atmosphere, and a protective magnetic field. In contrast, Venus' surface temperatures are the hottest in the Solar System (~460 ˚C), its atmosphere is the thickest of any rocky body and is rich in the greenhouse gas carbon dioxide, and it has no global magnetic field. Yet Earth and Venus are similar in size, with Venus only 30% closer to the Sun than Earth. In theory, this should make the surface temperature of Venus only 15˚C hotter than Earth.

What set Earth and Venus on such different paths? Is this a manifestation of their locations in our Solar System or did chance events during their formation play a critical role? Determination of Venus' atmospheric and surface compositions, as well as its geophysical properties, are needed to provide insight into whether Earth and Venus were more similar in the past and evolved differently through time (e.g., see Question 6 for discussion of their divergent climate evolution), or whether these neighboring planets were made differently from the start. For example, knowing the bulk and isotopic composition of Venus, which along with Earth, makes up >90% of the inner Solar System's mass, would enable the compositional heterogeneity of the inner Solar System to be much better understood, and would reveal Earth-Venus similarities or differences. This in turn, would provide crucial insights into Earth-Moon origin (Q3.3).

Unlike Earth, Venus has a slow retrograde rotation and lacks a moon. These differences, as well as Venus' current lack of a magnetic field, could reflect differences in the types of impacts experienced by Venus during the end of its accretion. The surface of Venus is thought to be young (~300–700 Ma) and the diameter of the largest crater is only 280 km, far smaller than those on Mars or other terrestrial bodies. Thus, Venus does not, and would not be expected to, display clear evidence of any giant or basin-forming impacts.

During the planetary accretion phase, relatively large planets, such as Earth and Venus, may establish a compositionally stratified structure, meaning that the outer part of the core is more enriched in light elements and therefore less dense. Such a structure is stable and would prohibit a core convection and therefore generation of a dynamo. If the stable structure is reset by a large impact, a dynamo generation is possible—this could have happened to Earth but not Venus. Although this could potentially explain why Venus does not have a moon or a magnetic field, it is not clear if Venus could have avoided giant impacts, given that simulations of terrestrial planet accretion find these are common for planets as massive as Earth or Venus.

Alternatively, some argue that Venus experienced two giant impacts (Alemi and Stevenson, 2006) September). The first generated a moon and provided Venus with a rapid prograde spin, while the second was oppositely oriented and caused Venus to rotate in the reverse direction. The latter caused the prior moon to tidally evolve inwards and eventually be lost. If the angular momenta delivered by the first and second impacts nearly cancelled out, the resulting spin angular velocity would have been small, potentially consistent with the long spin period of Venus (243 days). Alternatively, Venus' current spin state may reflect a combination of internal and atmospheric tides, core-mantle friction, and planetary perturbations. For an initial Venus day much shorter than its year, dissipation in the planet's interior due to solar tides is the dominant mechanism that causes the planet's spin rate to slow, and Venus' current state can be achieved if Venus' initial day was > 48-hr and its average tidal dissipation comparable to that of the Earth (e.g., Correia and Laskar 2001). Thus, whether Venus experienced a large impact or had a moon remains controversial. Observations to constrain the core structure and other interior layers of Venus would be valuable to gain insight into the planet's evolution and the role of giant impacts in its history.

Q3.4b. What was the nature of Mars's formation and how did its small moons originate?

Mars's small mass compared to those of Earth and Venus may be evidence that some process (e.g., giant planet migration; Q3.2) depleted material from its orbital region before it formed. Martian meteorites display large isotopic variations in now extinct systems (e.g., hafnium-tungsten and samarium-neodymium) compared to the much more limited isotopic compositions observed in lunar and terrestrial samples. Mars's tungsten, neodymium, as well as strontium, isotopic compositions nominally suggest that it accreted earlier than the assembly of the Earth and Moon, although these results are sensitive to the nature of core formation, late accretion, and details of the isotopic evolution models. Its small mass and inferred early formation time have led to the suggestion that Mars is a leftover planetary embryo that may not have experienced an extended phase of giant impacts like the other terrestrial planets. However, tungsten, and to a lesser extent neodymium and strontium, isotopic compositions do not appear to vary systematically with many geochemical characteristics of the samples indicating that our understanding of Mars' isotopic composition is incomplete. Additional samples from Mars are needed to more accurately decipher the process responsible, as well as the age of formation of Mars.

One possible proxy for giant impacts are moons. While Mercury and Venus have no moons, Mars has two small irregularly shaped moons that resemble primitive asteroids. Views on the origin of Phobos and Deimos have changed dramatically over the past decade. Originally, they were thought to be asteroids gravitationally captured intact by Mars. However, Phobos and Deimos have very small eccentricities and inclinations that are not easily explained by the capture model; such moons tend to have large eccentricities and inclinations, and inclinations cannot be sufficiently reduced by Mars's tidal forces.

The most attractive alternative scenario is that the moons accreted from a disk formed by a large impact, as is thought to have been the case for the Earth's Moon (e.g., Rosenblatt 2011; Citron et al. 2015; Canup and Salmon 2018). It is plausible that this collision was the same one that formed hypothesized Borealis basin, an approximately 10,600 by 8,500 km impact feature that may be responsible for Mars's crustal dichotomy between the northern lowlands and southern highlands (Andrews-Hanna et al. 2008), or by the impactor that produced the Utopia or Hellas basins. The implied impactor falls in the 600-km to 1000-km size range, proportionally much smaller relative to Mars than Theia compared to Earth. Although the full effects of a Vesta-to-Ceres sized body smashing into Mars have yet to be explored, it is difficult to imagine that martian core and mantle were unaffected. Models suggest that Phobos and Deimos may be the last survivors of multiple satellites originally formed by this impact, whereas other models suggest that Phobos may have formed more recently (Hesselbrock and Minton 2017; see Question 8). Other possibilities, such as disintegration from a single progenitor (Bagheri et al 2021), are speculative but remain in play. A critical step in constraining the origin of Phobos and Deimos will be to determine their bulk compositions and interior structures.

Q3.4c What conditions led to Mercury's anomalously high-density and large core?

Mercury's large core comprises some 69 to 77% of its mass (Hauck et al. 2013), as confirmed by gravity measurements by the MESSENGER mission. The origin of its anomalously large core is thought to be key to understanding the formation of Mercury and its subsequent thermochemical evolution. Numerous hypotheses for Mercury's origin have been proposed over the last several decades, but data returned by the MESSENGER mission have narrowed the likely formation mechanisms of Mercury into two competing groups (1) a mantle-stripping giant impact processes and (2) formation from highly-reduced metal-rich precursor materials from the inner portion of the protoplanetary disk.

There has long been a proposal that the anomalously high metal content of Mercury was produced by a high-velocity giant impact that blasted away most of Mercury's primordial mantle (e.g., Benz et al., 1988). However, MESSENGER data also show that Mercury's surface is rich in volatiles, such as sulfur, potassium, and chlorine (e.g., Nittler et al., 2011; Evans et al., 2015). These data have challenged the idea of forming the high-density planet by a disruptive impact, which is expected to have resulted in extensive volatile loss. In contrast, the idea of volatile depletion through giant impact processes for large planet-sized bodies has not been adequately demonstrated, and Mercury's volatile-rich nature may not disqualify the giant impact model (Ebel and Stewart, 2018). Alternatively, it has been suggested that Mercury might have been the smaller of two planets involved in a hit-and-run collision that stripped its mantle while avoiding disruption and extensive heating (Asphaug and Reufer 2014). Another possible explanation for Mercury's origin is that its large metallic core and highly reduced surface are an outcome of its accretion from highly reduced, metal-rich materials that differed from the building blocks of the other terrestrial planets by virtue of Mercury's innermost location in the Solar System. However, it is not clear to what extent Mercury's reduced nature can be attributed to impacts onto Mercury's precursor planetesimals or even Mercury itself.

Both of these formation mechanisms have different implications for the subsequent thermochemical evolution of Mercury, and both possibilities must be considered and weighed when interpreting Mercury's nature. Geochemical, mineralogical, and isotopic measurements of Mercury materials are needed to constrain the origin of Mercury, as are better constraints on the planet's interior structure.

Mercury's fractionally large core has interesting implications for exoplanets. Two innermost planets of Kepler-107 have large density variations, indicating that one of them is Earth-like and the other one is Mercury-like. This diversity could have been produced by an impact (Bonomo et al., 2019).

Understanding the origin of Mercury's core would give us key insights to planetary diversities in extrasolar systems as well.

Strategic Research to address Q3.4:

- **Determine the interior structure and bulk composition of Venus** by constraining the moment of inertia with ground-based telescopes, orbiters, and/or seismic measurements, and through atmospheric and surface observations.
- **Determine the isotopic compositions of Mercury and/or Venus** through remote sensing, in situ measurement, sample return, and/or identification of meteorites originating from these planets
- **Determine the origin of the Martian moons** by comparing chemical and isotopic ratios of Mars, Phobos, Deimos and asteroids and constraining their interior structures.
- **Determine the formation time of Mars** through isotopic analyses of diverse martian samples.
- **Determine the origin of the large core of Mercury** by in-situ geochemical, mineralogical, and isotopic measurements.

Q3.5 HOW AND WHEN DID THE TERRESTRIAL PLANETS AND MOON DIFFERENTIATE?

Present-day characteristics of all differentiated bodies are closely linked to the mechanisms and timing of their initial differentiation, which involved the formation of metallic cores, rock mantles, crusts, and in some cases volatile-rich atmospheres/hydrospheres. Significant research has been devoted to this topic over the last few decades, and open questions remain.

Q3.5a What were the mechanisms of primordial differentiation?

Differentiation is inextricably linked to heating during planetary formation, produced by the decay of radiogenic elements and the conversion of gravitational energy to heat during accretion. For planetesimals and small bodies, radiogenic heating dominates (Dodds et al., 2021). Two classes of radiogenic heat sources are short-lived nuclides with half-lives less than 1 million years (^{26}Al and ^{60}Fe), and long-lived radiogenic nuclides with half-lives greater than ~1 billion years (^{235}U, ^{238}U, ^{232}Th, ^{40}K). Short-lived heating led to differentiation on smaller bodies formed within a few million years of the oldest dated materials in the Solar System (calcium-aluminum inclusions, or CAIs), producing the parent bodies of iron meteorites, Vesta's iron core (Zuber et al, 2011), and partial differentiation on Ceres (Castillo-Rogez et al., 2019). Ancient ages, such as the 4565.4 ± 0.2 million year age determined for a meteorite sample from Vesta (Wadhwa et al., 2009), suggest that 500-1000 km class planetesimals were differentiated as they formed. Decay of long-lived radionuclides and accretional heating are thought to be the dominant heat sources for larger bodies.

Differentiation processes involving magma oceans are thought to be responsible for the structure of most large bodies in the inner Solar System, including the Moon, Earth, Mars, and asteroids such as Vesta (see also Question 5). This involves progressive crystallization of minerals: light minerals buoyantly rise to the surface, forming a primary crust, while denser minerals sink. Geologic activity on all of the terrestrial planets has all but destroyed the surface evidence of these primary crusts. However, this record is preserved at the Moon, where plagioclase (the first buoyant minerals that crystallize out of a magma ocean) rose to the surface to form the lunar highlands, providing the most compelling evidence for magma ocean differentiation. A giant impact can produce a global magma ocean, and large impacts may re-melt substantial fractions of a planet's surface, so that it may be more accurate to think of a series of magma oceans as opposed to a single one. Magma oceans likely survived for millions to tens of millions of years, interspersed by relatively temperate conditions. One notable exception is Jupiter's moon Io, which may maintain a long-lived magma ocean today tidal heating (Khurana et al. 2011; Question 8).

Although Venus is similar in size to Earth, it is unclear how or even if it underwent primordial differentiation. Venus may have experienced an ongoing differentiation process analogous to plate tectonics on Earth or some as of yet undefined differentiation process. One hypothesized mechanism involves a major crustal instability that rapidly resurfaced the Venusian crust (e.g., Strom et al., 1994). The other hypothesized mechanism involves progressive volcanism (e.g,. Smrekar et al., 2010). Currently, the surface resolution and crater statistics are unable to distinguish between the two hypotheses.

Improved determination of how the inner planets differentiated is critical to understanding their formation conditions and early evolution. Measuring noble gas concentrations in the Venusian atmosphere would provide an important constraint on degassing from the interior. Determining the ages and composition of the enigmatic tesserae regions on Venus would test the hypothesis that these large plateaus are analogous to continents, suggesting the presence of water when they formed (Gilmore et al., 2017). Another important question is what criteria, such as primary heat source, size and composition of the body, or efficiency of thermal blanketing by an atmosphere, dictate the style of planetary differentiation. Finally, better understanding the relationship between mechanisms of silicate differentiation and the resulting characteristics of planetary cores, mantles, and crusts on differentiated bodies is needed. For example, would some mechanisms lead to more sulfur within planetary cores, as has been suggested to explain the larger and lighter core of Mars relative to Earth? Could the retention or loss of volatile elements and compounds, including water, be related to planetary differentiation mechanisms? While it is thought that magma oceans are effective means for losing volatiles or generating transient or permanent atmospheres via outgassing, it is not clear if other differentiation mechanisms would lead to similar outcomes.

Q3.5b What was the timing and duration of primordial differentiation?

All of the inner planets differentiated, but whether they underwent this process at the same time remains unknown. Determining ages for planetary differentiation is challenging. Relative chronometers, such as crater counting, have uncertainties that stem from the limited availability of samples from planetary surfaces with which to calibrate crater densities with absolute ages. Various radiometric chronometers have underlying assumptions that may be valid only under certain circumstances, date different aspects of differentiation, and/or be disturbed by later geologic processes.

Great progress has been made in dating core formation using the Hf-W (hafnium-tungsten) isotopic system. An isotope of hafnium that is concentrated in silicate mantles and crusts (^{182}Hf) has a short half-life of 9 million years, and therefore only records the first 60 million years or so of Solar System history. Remarkably, ^{182}Hf decays to an isotope of tungsten that is strongly partitioned into metallic cores (^{182}W). A core that formed early will not have much ^{182}W, whereas the remnant mantle will be enriched in ^{182}W, compared to chondritic abundances. Thus, the Hf-W isotopic system can be used to obtain core formation ages from samples derived from both metal cores as well as silicate mantles and crusts. This system has revealed that iron meteorites represent pieces of planetesimal cores formed in the first 1-2 million years of the Solar System (Kruijer et al., 2014). Although complicated by the fact that late accretion can deliver tungsten to planetary mantles after core formation (see section Q3.5d), it has also been possible to demonstrate that the cores of Mars and Earth formed within the first 1–10 million years and 30–200 million years of Solar System history, respectively (Nimmo and Kleine, 2007; Nimmo and Kleine, 2015).

Primordial silicate differentiation has been dated on large bodies, such as the Moon, Earth, and Mars, using a variety of long and short-lived chronometers. This requires independent knowledge of how a sample, or groups of samples, formed. Most investigations presently assume the silicate portion of large planetary bodies differentiated from an isotopically homogeneous magma ocean at the end of planetary accretion. Isotopic measurements on the solidification products of these magma oceans then allows the age of differentiation to be determined. Using various approaches, the age range of lunar differentiation is reported to be 4.33 to 4.56 billion years old (see summary in Borg et al., 2015). This is a remarkably large range for a body that we have arguably sampled better than any other outside of Earth. A major outstanding issue is that lunar differentiation ages are based almost solely on samples collected by the Apollo missions, which we now know sampled an unusual terrane on the Moon (Jolliff et al., 2000). Thus we do not know if these ages represent planetary scale differentiation, or a widespread, but localized melting event that followed planetary differentiation.

The age of silicate differentiation of the Earth is better constrained to 4.38 to about 4.42 billion years ago (Caro, 2011; Valley et al., 2014). Variations in ^{182}W isotopic compositions have been found in ancient crustal rocks and modern oceanic lavas suggesting that differentiation of Earth may have isolated some silicate reservoirs very early on that have been preserved for billions of years until the present day. Ages for silicate differentiation on Mars based on shergottite meteorites are also quite tightly constrained at ~4.50 to 4.53 billion years ago, although not all samples (e.g., Nakhlite meteorites) yield concordant differentiation ages. Furthermore, martian meteorites have large variations in ^{182}W, that perhaps implies even earlier differentiation, or more extreme fractionation of the parent element (Hf) from the daughter element (W) during differentiation. While differentiation ages of planetesimals and asteroids have not been as well determined, the observation that basaltic samples from these bodies crystallized within a few million years of the beginning of the Solar System indicates differentiation was very early. Our knowledge of how these or other bodies cooled is not complete; cooling is typically defined by thermal modeling which yields different results depending on the assumptions used in the calculations. The Moon for example is estimated to cool (that is, experience freezing of its magma ocean) in anywhere from 2 to 200 million years (Elkins-Tanton et al., 2011; Maurice et al., 2020).

Ages for solidification of the lunar magma ocean have been addressed by determining isotopic ages on three types of lunar magma ocean crystallization products cumulates including: (1) mafic cumulate mare basalt sources, (2) crustal rocks of the ferroan anorthosite suite, and (3) late-stage crystallization products of the lunar magma ocean (i.e., urKREEP, material rich in potassium (K), rare-earth elements (REE), and phosphorus(P)). Detailed reviews of these ages, their merits, and their inconsistencies are provided in Papike et al. (2018), Nyquist et al. (2001), Borg et al. (2015). The majority of ages fall in the range of 4.30 to 4.38 billion years ago, and provide a lower age limit for accretion of the Moon and the giant impact. However, all of these age determinations are based on Apollo samples collected from the lunar nearside, and therefore could represent either lunar magma ocean solidification or a widespread, but regional, magmatic event focused on the nearside (e.g., Borg et al., 2011; Tartèse et al., 2019). The age range of lunar magma ocean cumulates is concordant with rocks of the Mg-suite, which are not currently thought to be products of lunar magma ocean solidification, suggesting the ages may not record lunar magma ocean solidification. Alternatively, the resolution of current chronometers may not be sufficient to distinguish two nearly contemporaneous, but geologically unrelated events. Finally, it should be noted that there are a few seemingly accurate ages that are older than 4.38 Ga (Tartèse et al., 2019; Borg et al., 2020). Additional high precision isotopic measurements are needed on crustal and mantle rocks collected outside the region sampled by the Apollo astronauts to unravel the chronology of the earliest geologic events in the history of the Earth-Moon system.

Overall, small bodies appear to have differentiated earlier than large bodies, perhaps reflecting a combination of the size of these bodies and their proclivity to lose heat, and the preponderance of ^{26}Al in the early Solar System that provided an important heat source during early times and stages of accretion. Large bodies, such as Mars, seem to have completed differentiation tens of millions of years later. Later differentiation in the Earth-Moon system was probably the result of a combination of stochastic events including the Moon-forming giant impact, and may have been influenced by early tidal interactions between Earth and the Moon. Due to the lack of samples, no differentiation ages can be estimated for other bodies like Mercury and Venus. Fundamental questions remain as to how the timing and style of primordial differentiation was linked to a body's size and the physical environment and chemical conditions in which it formed. Additional samples, more detailed thermal modeling, more thorough understanding of the surface geology and interior structure of differentiated bodies are required to make further progress.

Q3.5c What were the causes of variation in oxygen fugacity in differentiated planetary bodies?

A notable feature of the Earth is the difference in oxidation between its surface and interior. On a large scale, the core is metallic, whereas Earth's atmosphere is ~20% oxygen, requiring progressive oxidation over time. This intrinsic thermodynamic variable is known as oxygen fugacity, sometimes abbreviated to fO_2. Like Earth, Mars has the same general variation in fO_2 with a reduced core and an oxidized crust, giving the latter its moniker as the 'red planet'. In contrast, studies of meteorites from asteroids and lunar Apollo samples suggest that the range of oxidation state within these bodies is more restricted. Data from MESSENGER indicate that Mercury is the most reduced of the terrestrial planets, with an fO_2 well below most asteroids and the Moon, and it shows evidence of a crust that is more reduced than its interior, opposite to what is exhibited by Earth and Mars. Such differences may at least in part reflect differences in body size (which may affect iron chemistry and resulting redox state during primary differentiation) and volatile content.

Oxygen fugacity changes the geochemical behavior of elements in common geologic processes and therefore has a profound effect on their distribution with a planetary body. On Earth, subduction leads to oxidization of the mantle, changing the valence of elements like uranium (U) which is insoluble in the reduced state (U^{+4}) and soluble in the oxidized state (U^{+6}). This process has had a remarkable effect on Earth's surface and atmosphere. On Mercury, the fO_2 is sufficiently low that many elements that typically are partitioned into the crust during differentiation are instead partitioned into the iron, or iron-sulfur, rich core.

Despite its profound importance for constraining geochemical processes we do not fully understand the primary drivers of oxidation and reduction. Oxidation and reduction could result from biologic activity, as has been suggested for Earth, or abiotic, as appears to be the case for Mars. We also do not understand the relationship between oxidization conditions on the surface of a body and those in its interior. Further studies of the oxidation state of Mercury and Venus will be important to addressing such issues. For example, an oxidized crust on Venus, might reveal fundamental processes for how planetary differentiation can—and cannot—lead to life-sustaining environments.

Q3.5d How did late accretion affect planet composition and chemistry?

Formation of metallic cores had a profound role in shaping the chemistry of planets. In addition to removing iron, core formation would have removed precious metals like platinum or gold from the silicate portions of planets almost entirely. That these siderophile (iron favoring) elements exist in the Earth's crust today is attributed to what is termed "late accretion", the addition of material rich in precious metals and volatile compounds to the Earth's upper layers by impacts after core formation had essentially ended (Figure 3.2). For Earth, the precious metal inventory in the crust and mantle can be accounted for by the addition of only about 0.5% to a few percent of the Earth's mass (e.g., Kleine and Walker, 2017; Marchi et al. 2018). Analysis of martian and lunar samples demonstrate that Mars experienced a similar amount of late accretion, whereas the Earth's Moon experienced a far smaller proportion.

Understanding why different bodies in the inner solar system appear to have different proportions of late accreted materials is critically important because it provides insights into the planet formation process. Various arguments now suggest that late accretion was dominated in mass by large, up to lunar-sized projectiles leftover from planet accretion (e.g., Bottke et al. 2010; Pahlevan and Morbidelli 2015; Brasser et al. 2016; Marchi et al. 2018). Therefore, understanding the size distribution of late accretion projectiles, and how the addition of large differentiated projectiles affect planetary surface composition and chemistry is of great importance. Addition of cometary material with its copious amounts of ices would be substantially different from accreting a volatile element depleted body like our own Moon, for example. Likewise, we do not know if the late accreting materials have, on average, fundamentally different compositions from the materials responsible for the earlier main phases of accretion. Such compositionally heterogeneous accretion has been used to explain the volatile content of Earth and the formation of its oceans and seems generally consistent with dynamical models that suggest that late accreting material would increasingly have been derived from the outer Solar System. Finally, determining how water and other key volatile elements (e.g., carbon, hydrogen, oxygen, phosphorus, and sulfur) were supplied is critical for our understanding of what makes a habitable planet. Means for addressing these issues lie in obtaining a refined understanding of late accretion from study of available materials linked with improved modeling efforts. Determining late accretion components to Venus and Mercury, currently unconstrained, are particularly important needed data.

Strategic Research to address Q3.5:

- **Determine the age relations between the oldest lunar crustal and mantle rocks** by dating lunar crustal and mantle rocks, which may be potentially found at the South-Pole Aitken basin.
- **Reveal the mechanisms of planetary differentiation on Venus** by measuring ages and composition of the tesserae regions with spacecraft observations or sample return.
- **Determine the age and duration of primordial differentiation on Earth, Moon and Mars** through isotopic analysis of samples collected from new locations on Moon and Mars, and through thermal modeling.
- **Determine and compare the mechanisms of differentiation, size of body, and location of bodies in the Solar System** through sample analysis, spacecraft observation, and geochemical/geophysical modeling.
- **Determine the oxidation state of planetary surfaces to understand the primary drivers of redox conditions** with spacecraft observations and sample analysis.
- **Determine the contribution of outer Solar System materials to the inventory of the inner Solar System planets** through measurements of the volatiles and refractory components of water-rich asteroids and comets by telescopic observations, in situ measurements, and/or analysis of returned samples.
- **Assess the contribution and effects of late accretion on the post-differentiation inner planets** by analysis of ancient terrestrial materials, samples from regions of Mars and the Moon likely to be derived from each world's mantle, and/or samples derived from Venus or Mercury, and through improved dynamical and geochemical modeling.

Q3.6 WHAT ESTABLISHED THE PRIMORDIAL INVENTORIES OF VOLATILE ELEMENTS AND COMPOUNDS IN THE INNER SOLAR SYSTEM?

Volatile elements include a wide range of geologically important elements, including hydrogen, fluorine, chlorine, nitrogen, carbon, and sulfur, as well as noble gases and alkali elements. These elements encompass nearly all of organic chemistry and water, including the life-essential elements. Although both differentiated and undifferentiated rocky planetary bodies in the inner solar system have nearly chondritic abundance ratios of refractory elements typically partitioned into silicate minerals (lithophile elements), differentiated rocky bodies are depleted in volatile elements compared to undifferentiated rocky parent bodies or CC meteorites. The magnitude of volatile depletion seems to correlate very roughly with the 50% nebular condensation temperature for a given element, with those with lower condensation temperatures depleted more strongly.

There are a wide range of explanations for the origin and timing of observed volatile depletions and isotopic compositions. For example, some models suggest that depletion occurred prior to the assembly of the rocky parent bodies from which samples derive. In contrast, other models call for parent body processes to account for the loss of volatiles. The timing of water delivery is one of the most researched topics, and there is still an unsettled debate as to whether water was inherited from the nebula, delivered during the primary stage of accretion, or delivered during the waning stages of late accretion. In addition, the relatively high abundances of water in Earth, relative to the Moon, suggests that processes other than location in the Solar System are responsible for producing the volatile element inventories of the terrestrial planets. The timing of planetary volatile depletions, and how volatile elements were successfully retained or added to planets, like Earth, remains key issues associated with understanding how, when, and why life began.

Q3.6a What were the primordial sources of volatiles?

By determining the chemical and isotopic makeup of elements like hydrogen, carbon, nitrogen, and oxygen in a variety of materials, including the Sun, atmospheres, planets, moons and primitive bodies, several key end-member compositions have been identified in the Solar System.

The first end-member is the **solar** composition, which represents that of the early proto-solar nebula, the cloud of ~99% gas (mainly H_2, CO, N_2, and noble gases, like helium or argon) and ~1% dust from which the Solar System emerged. Even some rocky planets have compositions approaching this endmember, including Earth. How this composition was obtained is not fully resolved. In one model, if the planets formed early enough, solar nebula gas could have been gravitationally trapped into primary atmospheres and eventually dissolved into planet interiors (Porcelli et al., 2001). Another model suggests that solar ions could have also been implanted from solar wind irradiation onto dust and/or planetary surfaces later incorporated into the planets (Vogt et al., 2019).

The second is a **chondritic** component. This endmember is defined from analysis of primitive meteorites (chondrites), in which volatile elements are present either as hydrated minerals or as organic material hosting hydrogen, carbon, and nitrogen, or both. Primitive meteorites which originate from leftover asteroids are our best representatives of early-formed planetesimals that contributed to the formation of inner planets. The non-carbonaceous chondrites (**NC**) are generally (but not always) volatile-poor and might have originated from a region in the disk located between the tar line (a radial distance from the Sun where organics could have survived) and the snow line (a radial distance beyond which water could exist as ice). Carbonaceous chondrites (**CC**) can contain up ~10% equivalent water in hydrated minerals and a few percent carbon and nitrogen as organics. Given their generally more abundant water, CC are thought to have formed beyond the snow line and either more distant from the Sun than NC, and/or at a different time in the evolution of the disk.

The third component is **cometary.** Comets are our best representatives of planetary bodies formed beyond the original orbits of the giant planets, and are 10–50% water and other compound ices by mass. Given that their trajectories can cross the orbits of the terrestrial planets, they are often regarded as potential sources of inner planet volatiles.

The interplay of these different contributors shaped the volatile inventories of the inner planets. Despite advances in knowledge of what the sources of volatiles were, how these three major reservoirs were established in detail, and what the ultimate interstellar source of these volatiles was, is largely unknown. Analysis of volatile elements from samples returned from asteroids, Mars and Moon, volatile-rich deposits on the Moon potentially representing cometary compositions, and ultimately comets will provide the principal means of making further progress on these questions through comparison with compositions determined for planetary atmospheres measured in situ and on returned sample.

Q3.6b How was the inner Solar System populated with volatiles and how did volatile delivery evolve with time?

Hydrogen, carbon, nitrogen and oxygen were likely incorporated into growing planetesimals in the form of ice and organic dust after the Solar System had cooled following its initial formation. Volatile-rich dust was inherited from the interstellar medium (ISM) or synthesized by photon-gas interactions in the parent molecular cloud or at the protoplanetary disk surface. Importantly, hydrogen, carbon, nitrogen, and oxygen trapped in chondrites, comets, and the atmospheres of the terrestrial planets have elemental and isotopic compositions markedly different from those of the protoplanetary disk as determined from the ISM or local gaseous reservoirs such as the giant planets, precluding a direct genetic relationship between the building blocks that make up the terrestrial planets and the gaseous solar nebula

Gas and dust were processed and distributed throughout the protosolar nebula by turbulent mixing, with (1) preferential preservation of molecular cloud signatures in the colder regions of the outer Solar System, and (2) organo-synthesis mainly occurring within the irradiated regions of the protosolar nebula. The spatial distribution of hydrogen, carbon, nitrogen, and oxygen was then mainly controlled by the location of the tar and snow lines. As discussed in Questions 2, 3.1 and 3.2, planetesimal/planet formation and giant planet migration led to the scattering of outer Solar System planetesimals, some of which entered the terrestrial planet region and were captured inside the asteroid belt.

The Earth is thought to have accreted mainly from dry material of the NC type. Volatile elements contributing wetter CC materials, namely planetesimals, comets, and dust from the giant planet zone, were accreted later as these materials became dynamically accessible. This compositional shift is consistent with the isotopic compositions of terrestrial planet noble gases as well as hydrogen and nitrogen stable isotopes observed in the mantle and atmosphere, respectively. Although this model is appealing, one inconsistency is that deuterium-to-hydrogen (D/H) ratios determined for a significant subset of comets are about twice as high as D/H values measured on Earth. To what extent this scenario is applicable to the other terrestrial planets is unknown because representative samples of planetary interiors are limited so far to Earth. Likewise, the impact of the early Sun's activity onto terrestrial planet atmospheres cannot be assessed without detailed knowledge of the atmospheres of Mercury and Venus. These constraints can be used by numerical simulations of planet formation to test which groups of bodies and processes are the most important for volatile delivery.

The timing of volatile delivery to planets remains poorly known, despite having a framework chronology of Solar System events derived from meteorites. There is evidence, however, that both Earth and Mars incorporated nebula gas during their differentiation from the noble gas compositions trapped within terrestrial rocks and martian meteorites, respectively. Overall, it seems likely that the initial delivery of volatiles into the inner Solar System may have occurred when accretion was taking place. Limited volatile delivery from impacting carbonaceous chondrite asteroids and comets continues to the present day (discussed in Questions 4, 5, and6) indicating volatile elements have been added to the terrestrial planets by this process as well. Further research should constrain the characteristics and abundances of nebular gas added to planetary interiors during initial accretion. In addition, developing broad and geochemically consistent models to understand potential late addition of CC materials to planetary bodies should be developed. Polar volatiles in the permanently shadowed regions of the Moon and Mercury may also retain a record of volatile delivery, although the exact age and origin of these volatiles is debated (Question 6).

Q3.6c How were primary volatile-rich reservoirs (atmospheres and oceans) produced?

Of the terrestrial planets, only Mercury does not have a substantial, visible atmosphere. The present-day atmospheres of Venus and Mars have comparable compositions, namely >90% CO_2 and a few percent N_2, but they are at substantially different pressures and have different water contents. Venus might have had early oceans that subsequently vanished. Geological evidence on Mars attest to the presence of oceans within the first few hundreds of million years.

Noble gas abundance patterns of the three planets are comparable, but this is difficult to reconcile with different atmospheric escape processing otherwise suggested by isotope variations. All three planetary atmospheres show a depletion of xenon relative to lighter noble gases. On Earth, this xenon depletion is often accounted for by prolonged hydrogen escape from photodissociation of H_2O. This could also be the case for Mars but at different periods of time. The case of Venus is unknown due to the lack of related atmospheric measurements.

Earth's atmosphere has been strongly modified by biological processes, and the atmospheres of Venus and Mars have been modified by runaway greenhouse effects or extensive mass loss, respectively. Despite eradication of the primary volatile-rich reservoirs on these bodies, evidence can still be found for these early stages from volcanic rocks arising from deep mantle sources in Earth and from Mars (as meteorites). A major unanswered question is why Earth has solar compositions of light noble gases like helium and nitrogen, but chondritic abundances of heavier noble gases like krypton and xenon, whereas for Mars, all appear solar. This effect may relate to late accretion of NC or CC materials, but exactly how is unclear.

The isotopic composition of cometary noble gases measured on comet 67P/Churyumov-Gerasimenko by the Rosetta probe indicates that Earth's atmosphere, but not its mantle, has a substantial contribution of cometary materials. It is thought these additions occurred after the Moon-forming impact. The timing and extent of cometary contribution to the terrestrial planets and Moon is a fundamental problem that will require precise documentation of the compositions of inner planet reservoirs (mantle and atmosphere of Mars and Venus), as well as those of cometary materials returned to Earth.

The elemental and isotopic compositions of terrestrial noble gases indicate that the young Earth experienced several episodes of degassing and volatile loss during the first tens to hundreds of million years. In particular, extant and extinct gas isotopes with variable half-lives permit reconstruction of the degassing/tectonic history of Earth: early isolation (within 100 Ma) of the deep and the convective mantle regions, intense degassing within a few tens to hundreds million years, secular evolution of convection and crustal growth around 3 billion years ago, recycling of surface volatiles including water becoming effective during the last 2 billion years.

Mars might have experienced a different volatile history. Early oceans seem to have vanished within the first few hundreds of million years of its history, while its tectonic activity might have been too limited to permit replenishment of the atmosphere. Reconstructing the history of Martian volatiles will await the return of Martian sediments and igneous rocks. The volatile history of Venus is for the most part unknown and must await a new generation of in situ missions to that challenging environment.

Strategic Research to address Q3.6:

- **Determine in situ the origin, degassing history, and mantle-atmosphere exchange rate of Venusian volatile elements** by measuring the noble gas elemental and isotopic compositions of Venusian atmosphere, as well as the abundances and stable isotope compositions of H, C, N, O, S species.
- **Precisely determine the elemental and isotopic compositions of Martian mantle and atmospheric volatiles at present and in the past** by analyzing carefully selected atmospheric and solid samples with different ages and provenances returned from Mars.
- **Measure the abundances and isotopic compositions of volatile elements in asteroids from different radial locations and in comets** by analysis of returned samples and/or in situ investigations.
- **Determine the origin and abundances of volatiles in inner Solar system bodies** by conducting geochemical, petrologic, and spectral measurements of these bodies and their associated samples and by coupling results from planetary accretion models, laboratory experiments on volatile behavior, observations of volatile distribution in the asteroid and comet populations, and geochemical measurements from a wide range of parent bodies.
- **Study the behavior and the elemental/isotopic fractionation of volatiles in temperature, pressure, chemistry, and ionization conditions relevant to the formation of planetesimals, embryos, Earth and Venus** using laboratory experiments and modeling.

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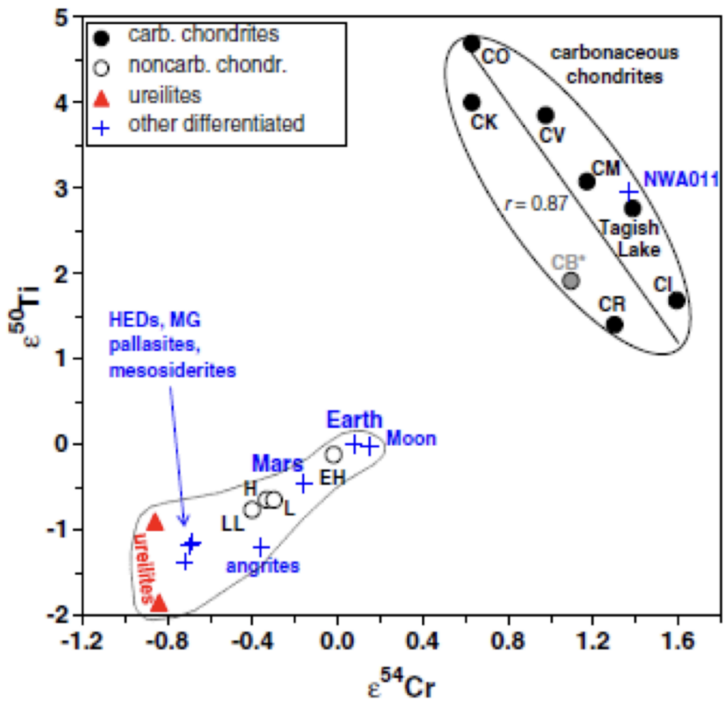
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