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Review article

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The accretion of planet Earth

Alex N. Halliday[1,3] ✉ & Robin M. Canup[2,3]

Sections

Abstract

Earth's origins are challenging to elucidate, given the lack of surviving terrestrial geology from the first 500 Myr of the Solar System. In this Review, we discuss breakthroughs in geochemistry and theoretical modelling that have advanced understanding of Earth accretion. Theory holds that solar nebula dust particles stuck together to form pebbles, concentrations of which gravitationally collapsed into ~100-km-sized planetesimals, which in turn accreted to yield planets. Isotopic variations in meteorites indicate that pebbles formed within the first 100 kyr of the Solar System, planetesimals melted and differentiated within a few 100 kyr, and Mars accreted quickly within 5 Myr. Earth's growth was more protracted, with >98% of its mass being accreted by the time of the Moon-forming Giant Impact at ~70–120 Myr. Earth is more enriched in *s*-process nuclides than chondritic meteorites, with a chemical composition affected by condensation, melting and loss. Early volatiles acquired from the nebula largely escaped, with the remnant volatiles being diluted by main-stage Earth accretion, accompanied by loss of nitrogen to the core and/or space. Areas for further research should include assessing mixing during large collisions and investigating the origin of very early mantle isotopic heterogeneities, which might indicate mass transfer from core to mantle over time.

[1]The Earth Institute, Columbia Climate School, Columbia University, New York, NY, USA. [2]Solar System Science and Exploration Division, Southwest Research Institute, Boulder, CO, USA. [3]These authors contributed equally: Alex N. Halliday, Robin M. Canup. ✉e-mail: alexander.halliday@columbia.edu

Key points

- Terrestrial planet accretion commenced with disk grains and high-temperature condensates sticking together to form pebbles, which in turn gravitationally coalesced to form planetesimals up to hundreds of kilometres in size. Planetesimals with metallic cores, sampled today as iron meteorites, were present within the first million years of the Solar System.

- Planetesimals collided to form Moon-to-Mars-sized planetary embryos in the presence of the solar nebula. Nebular dispersal triggered an era of giant collisions among the embryos that established the inner Solar System's architecture and, for Earth, culminated in the Giant Impact that produced the Moon.

- Although most of Earth's nucleosynthetic makeup is closest to that of enstatite chondrites, earlier (<50% by mass) stages of accretion had an isotopic signature intermediate between enstatite and ordinary chondrites. However, Earth is more enriched in those nuclides formed by slow addition of neutrons in large stars compared with all meteorites, and is different chemically from chondrites, particularly enstatites.

- These chemical differences partly reflect early melting and condensation in the disk, which produced fractionated chemical and isotopic compositions, but also result from subsequent losses and additions, especially of volatile elements, during accretion.

- Most lunar origin models fail to provide a natural explanation for the identical isotopic composition of the bulk silicate Earth and Moon for non-volatile elements. This isotopic match is particularly problematic for tungsten, which is sensitive to the nature and timing of core formation and is unlikely to result from the Giant Impact unless there was post-impact mixing and isotopic equilibration between the silicate Earth and Moon.

- The discovery of mantle isotopic heterogeneities generated in the first 100 million years of Earth's history has changed thinking on preservation of primordial reservoirs in the deep Earth, as well as the nature of Earth's late veneer, which could partially reflect a long history of compositional fluxes from Earth's core.

Introduction

Earth's present-day composition, including its water and other compounds essential for life, is inherited from accretion and fractionation processes that occurred during its formation and early history. By exploring the composition of Earth and other extraterrestrial bodies, as well studying processes occurring in other protoplanetary disks and planetary systems hundreds to thousands of light years away, Earth's planetary evolution can be unravelled with increasing detail, all the way from solar nebular dust to a segregated and dynamic planet replete with life (Fig. 1).

Most modern thinking on terrestrial planet formation followed the Apollo 11 to 17 missions (1969–1972). Dynamical simulations were developed that showed that terrestrial planet formation lasted tens of millions of years[1], culminating in violent planetary collisions like the Moon-forming Giant Impact[2]. In parallel, isotopic analyses of lunar and meteoritic samples (Box 1) allowed scientists to date key events in the earliest Solar System, providing a test of accretion models[3] and a fingerprint for the origins and provenance of materials in the young Sun's nebular disk[4,5]. Beginning in the 1980s, ground-based and space-based telescopes revealed gas-rich protoplanetary disks within which Sun-like stars form, allowing early accretion processes to be studied. Multiple observational techniques have now led to the detection of thousands of exoplanets, and the surprising diversity of exoplanetary systems implies that our Solar System structure is not typical and probably reflects specific, stochastic events[6]. Understanding of exodisk composition and morphology, as well as the physical properties of individual exoplanets and their atmospheres, will be further revolutionized by data from the James Webb Space Telescope.

This Review is motivated by the need to summarize the latest discoveries and thinking pertaining to Earth's origins. We discuss recent theories and evidence concerning the mechanisms and timescales for terrestrial planet accretion in our Solar System. We review advances in understanding the provenance of the main components from which Earth formed, based on isotopic compositions, and the modifications of these materials before, during and soon after accretion. We discuss current theories for the origin of Earth's Moon via the Giant Impact, and touch upon debates about the accretion of Earth's water and other volatile elements, which contributed critically to it being a successful harbour of life. We finish with one of the most intriguing latest discoveries — the preservation of extremely early mantle isotopic heterogeneities and what they reveal about the first ~100 Myr of Earth's history.

Dust to planetesimals

Stars form from the cores of molecular clouds that collapse because of self-gravity, in some cases triggered by supernova-produced shock waves. As a core collapses, gas and entrained dust grains containing too much angular momentum to fall directly onto the central protostar instead flow into a circumstellar disk that is rich in hydrogen and helium gas. The early Sun was surrounded by such a gaseous nebula for the first few to 10 Myr of its life, based on observations of young star–disk systems[7], until the gas was dissipated by stellar winds and irradiation by the Sun and nearby stars. This section describes how and when dust grains in the solar nebula evolved to produce the parent bodies of meteorites and the building blocks of planets.

Physical mechanisms

Collisions between initially micrometre-sized grains are thought to have led to the growth of millimetre-to-decimetre-sized pebbles, thanks to the combined effects of energy loss in inelastic, compacting collisions and surface sticking forces[8,9]. The processes responsible for growth from pebbles to ~100-km-sized planetesimals have been widely debated, because across this size range, sticking forces are negligible and interactions with the gaseous disk cause impact speeds during two-body collisions to be too high to allow for gravitational aggregation.

A key advance has been the identification of a planetesimal growth mechanism known as the streaming instability[10–12], which can occur when nebular gas and solid particles coexist. The gas is pressure-supported, and so it orbits the Sun at a somewhat slower azimuthal velocity than would a solid object orbiting at the same distance from the Sun. Orbiting particles experience a headwind as they encounter the slower orbiting gas, causing the particle orbits to lose energy and drift radially inward. If as particles drift inward a local concentration of particles forms, this concentration will accelerate the local gas

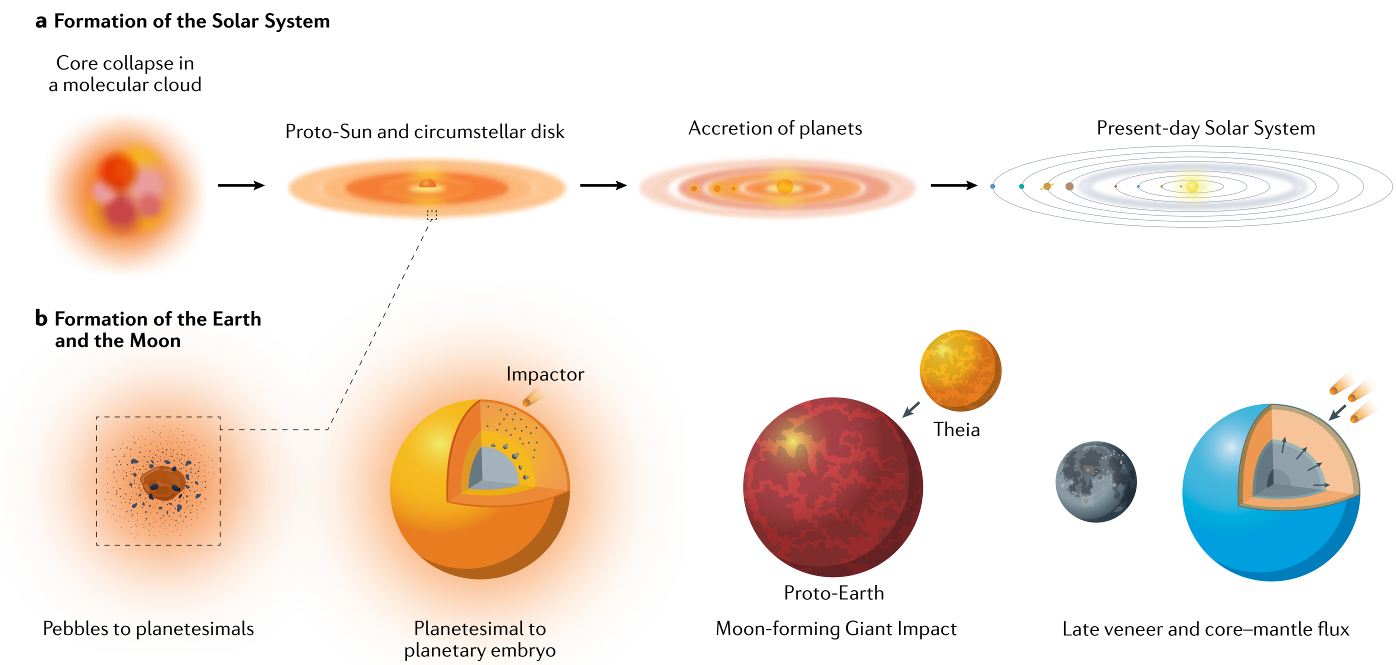


Fig. 1 | Formation of the Solar System, Earth and Moon. a, The Solar System formed from the gravitational collapse of a molecular nebula cloud, rich in hydrogen and helium gas, into a circumstellar disk with the proto-Sun at its centre. Within the disk, dust particles accreted to eventually form the planets. **b**, Collisions between initially micrometre-sized dust grains are thought to have led to the growth of millimetre-to-decimetre-sized pebbles, which in turn accreted into ~100-km-sized planetesimals. Planetesimals further accreted into ~1,000-km-sized planetary embryos, some with segregated cores. As the nebula dissipated, gravitational interactions between planetary embryos caused their orbits to cross, leading to protoplanet collisions. In the case of Earth, it is thought that a Mars-sized protoplanet called Theia collided with Earth (often termed the Giant Impact), and the resulting impact ejecta formed the Moon. After the Giant Impact, a small portion (<0.5%) of Earth's mass has been argued to be delivered by a bombardment of much smaller impactors, called the late veneer. However, the evidence for the late veneer is unclear, and the isotopic evidence could be explained partially by a flux of deep mantle and/or core components into the mantle over geological time. Parts of panel **b** adapted from ref.[232], Springer Nature Limited.

somewhat, lessening the strength of the head wind and the rate of the concentration's inward drift. The concentration can then continue to grow by accreting outer particles that are drifting inward more rapidly, and as the concentration grows, its effects on the gas strengthen, allowing its drift to slow further and its growth to continue. This positive feedback leads to an increase in the spatial density of solid particles, which can be sufficient to allow an entire region to rapidly collapse gravitationally into a 100-km-scale planetesimal[12,13]. Evidence supportive of planetesimal formation via streaming instability includes the properties of Kuiper belt binaries[14], and the inferred initial size distributions of the asteroid and Kuiper belts[12].

Meteoritic and observational constraints

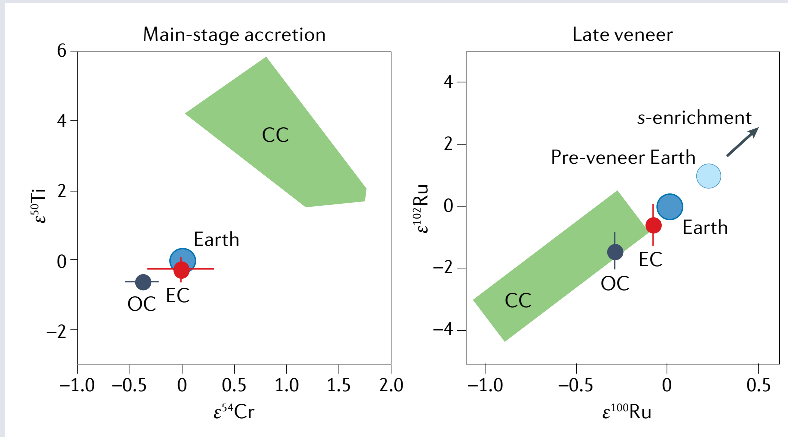
Meteorites (Box 1) provide an archive of conditions within the circumstellar disk and carry isotopic variations that constrain the timing of planetary accretion (Box 2). The best representatives of the nebular disk's composition (without its hydrogen and helium components) are chondrites, the commonest meteorites to fall on Earth. Inclusions rich in highly refractory elements (Ca- and Al-rich inclusions, CAIs) found in chondrites are the oldest macroscopic objects yet dated in the Solar System, setting its age to 4.56730 ± 0.00016 Ga (ref.[15]) (Fig. 2, Box 2). CAIs are thought to represent condensates, solidified partial melts and/or agglomerates that began forming within the first 10 kyr and that continued to form or be reprocessed over a few hundred thousand years[16,17]. Chondrules – small spherules in chondrites – have ages that span ≥3 Myr and so largely formed after CAIs. However, some chondrule ages overlap with those of CAIs, indicating that both classes of particles began forming in the very early disk[15,18]. CAIs and/or chondrules could represent pebbles (or constituents of pebbles) described above.

Observations of young stellar objects and their disks also provide insights into early accretion. Estimates of the total mass in subcentimetre-sized particles in 1- to 3-Myr-old disks[19] appear too low to explain observed exoplanet system masses, suggesting that most solids have already accreted into larger, observationally hidden sizes by this time[20]. Some observational evidence suggests the earlier growth of millimetre-sized pebbles in the collapsing envelopes of 100-kyr-old protostars[21,22].

The ages of the first large bodies in the Solar System with interiors differentiated into metallic cores overlain by lower-density silicate rock can be deduced using ^{182}Hf–^{182}W chronometry (Box 2). This chronometer has been applied to iron meteorites to obtain precise timescales for accretion of their differentiated parent bodies whose sizes, while uncertain, could be similar to the planetesimals discussed above[23–25]. With increasingly accurate methods of correcting for cosmogenic nuclides, it is apparent that accretion of some planetesimals occurred between 0.1 and 0.3 Myr after the formation of CAIs[4,26] (Fig. 2).

Box 1

Chondrites and the disk from which Earth formed



Meteorites provide crucial information on the mechanisms by which Earth formed, allowing its net accreted chemical composition, the precise timings of disk processes and planetary accretion, and the provenance in the disk of Earth's constituents to be determined.

With the exception of those from Mars and the Moon, meteorites come from asteroidal objects typically <100 km in size. Chondrites are from objects that never fully melted and have a composition similar to the Sun, ignoring highly volatile elements such as He. They include a fine-grained matrix that hosts presolar dust grains formed in earlier stars. Chondrites can be divided into carbonaceous — enriched in carbon, hydrogen and volatile compounds, and thought to come from further out in the Asteroid Belt and beyond; ordinary — the commonest to fall on Earth, lacking carbonaceous phases and with metallic components; and enstatite — similar but rare and highly reduced, with chemical affinities with the planet Mercury.

Refractory element (Ca, Al)-rich inclusions, or CAIs, are the oldest yet identified materials that formed within the Solar System (Fig. 2, Box 2). Also common are silicate-rich 0.1- to 1-mm spherules (chondrules), which, along with CAIs, could represent examples of early-formed small pebbles discussed in the text. However, some might be a by-product of planet formation.

Oxygen isotope heterogeneities across meteoritic classes are thought to reflect ultraviolet irradiation and photochemistry phenomena in the early nebular disk. Reconstruction of Earth's oxygen isotope composition from chondrites yields a mixture chemically unlike Earth. The same is true of nucleosynthetic heterogeneities. For the purposes of understanding Earth's accretionary makeup, the key isotopes of elements heavier than Fe (such as Ru) are those associated with the *s* (or slow) versus the *r* (or rapid) neutron-enriching nucleosynthetic process (see figure).

Each isotope is the product of a mix of these pathways, and each is shown as a deviation from a ratio. So, for example, ε^{100}Ru is the deviation in parts per 10,000 in the ratio ^{100}Ru/^{101}Ru relative to some standard value for Earth. Both neutron-enriching pathways can only be formed in massive stars. Carbonaceous chondrites (CC; green fields on figure), probably from colder, more distal parts of the nebular disk, are slightly enriched in *r*-process nuclides. The silicate Earth (dark blue circle) is richer in *s*-process components than any chondrites, the closest match being with the enstatite chondrites (EC; red circle) and ordinary chondrites (OC; grey circle), which are thought to have a provenance closer to Earth (based on figures in refs.[104,105]). The inferred pre-veneer Earth (light blue circle, right plot) had Ru-isotope compositions distinctly more enriched in *s*-process components than any chondrites[105]. Therefore, although the closest meteorite analogues to Earth in nucleosynthetic terms are enstatite chondrites, it is clear that some of the material accreted was more enriched in *s*-process components than anything in our meteorite collections.

Planetesimals to planet Earth

Once planetesimals formed, pairwise collisions between them – or between planetesimals and pebbles[12] – led to growth of 1,000-km-scale planetary embryos. After the solar nebula dispersed, giant impacts and mergers among the embryos continued until a few well-spaced and dynamically stable inner planets remained. Analyses of Earth's elemental and isotopic composition, in particular in comparison to those of the Sun and meteorites, provide crucial constraints on the nature and timing of these main phases of Earth's accretion, which established the planet's initial state and chemical inventories.

A multitude of planetary embryos

For planetesimal and larger-sized bodies, impact speeds during two-body collisions could be only modestly larger than the escape velocity, so that inelastic collisions can result in mergers. Once a body in an annular region of the disk becomes more massive than its local neighbours, it grows much more quickly and becomes the dominant body in that region. The dominant bodies that form in adjacent regions tend to be similar in mass, a pattern known as oligarchic growth. In the inner Solar System, oligarchic growth is predicted to yield a distribution of Moon-to-Mars-sized planetary embryos, having nearly circular and coplanar orbits, within ~1 Myr (refs.[27–29]). The system of many mini-planets would have been dynamically stable so long as the nebula was present, owing to interactions with the gas that keep orbital eccentricities low[30].

The predicted formation time for embryo-class objects generally agrees with meteoritic constraints. Isotopic analyses of the Steinbach iron meteorite suggest that its ~1,000-km-diameter parent body accreted and differentiated between about 1.3 and 1.8 Myr after CAIs[31]. For most silicate-rich igneous meteorites (achondrites), mainly derived from the crust and shallow mantle of differentiated bodies, the timing of primary accretion is harder to define because the parent to daughter ratios fractionated in subsequent silicate melting. Nonetheless, timescales of up to 5 Myr for accretion and core formation of some achondritic parent bodies have been inferred[32,33]. Mars — whose mass is about 10% of that of Earth and which could itself be a leftover planetary embryo — accreted in <5 Myr, based on ^{182}Hf–^{182}W chronometry[34] (Fig. 2). Embryos thus formed in the inner Solar System in the presence of the gas nebula, which persisted for a few to 5 Myr (Fig. 2). Earth accretion almost certainly started early too, but, as will be explained, took longer to complete.

The existence of the giant planets, which contain substantial H and He gas, implies that massive ice–rock cores capable of accreting gas had formed in the outer Solar System before the solar nebula dissipated. Formation timescales for Jupiter and Saturn of a few to 6 Myr have also been proposed, based on ^{182}Hf–^{182}W chronometry of early iron meteorites and the Solar System's nucleosynthetic variations[4], and constraints on gas giant satellite melting associated with early radiogenic heating, mainly by ^{26}Al[35,36] (Box 2, Fig. 2). Thus, at the end of the solar nebula's lifetime, the outer giant planets were present, whereas terrestrial planet assembly was still incomplete.

Box 2

Early Solar System chronometry

Isotope chronometers use naturally occurring radioactive decay to determine early Solar System timescales. They divide into:

- Those with a half-life >0.1 Gyr, such that the primordial radioactive parent is still measurable. For example, ^{235}U decays to ^{207}Pb with a half-life of 704 Myr. The Solar System and nearly all rocks are most reliably and precisely dated with the U–Pb system, providing an absolute determination of geological time.
- Short-lived chronometers with a half-life <0.1 Gyr, such that the parent is extinct. In this case, the former parent isotope abundance is determined from the relationship between parent and daughter element ratios and daughter element isotopic abundances among objects of the same age. Short-lived chronometers permit precise time resolution of early accretion processes.

Two short-lived chronometers are especially relevant to Earth's accretion:

- ^{26}Al decays to ^{26}Mg (half-life = 0.72 Myr), which is ideal for determining how quickly very early objects such as CAIs, chondrules and millimetre- to decimetre-sized pebbles formed. Its decay energy also provided a source of heat to achieve early melting.
- ^{182}Hf decays to ^{182}W (half-life = 8.9 Myr), whose characterization was aided by the development of new mass spectrometry techniques to measure isotopic abundances in trace amounts of tungsten. The Hf–W system provides a way to measure the rates of formation and early differentiation of planets. In principle, an entire planetary object and its first differentiation can be dated. Both Hf and W are refractory elements, such that a total planet's Hf/W ratio can be well established as solar (chondritic). Unlike the parent element (Hf), the daughter (W) is moderately siderophile (metal-loving) so the majority would be sequestered into the core during its formation. Such planetary fractionation of Hf from W provides insight into the timing of accretion and core formation because the short half-life generates W isotopic variations in proportion to Hf/W in the first 70 Myr, of the same order as the timescale of terrestrial accretion.

The era of giant impacts

As the nebula dissipated, gravitational interactions among terrestrial embryos caused their orbits to become eccentric, leading to crossing orbits and giant embryo–embryo collisions[37,38]. Since the late 1990s, *N*-body simulations that track this evolution[39,40] have shown that a final system of a few inner terrestrial planets typically results after ~100 Myr. The overall process is chaotic: simulations with only slightly different initial conditions yield different final planet positions, spacings and masses. Yet by performing many such simulations, a statistical range of possible outcomes is revealed that can be used to infer conditions best able to replicate the inner planets.

An important open issue is how the early orbital evolution of the giant planets might have affected inner planet accretion. The orbits of Pluto and other Kuiper Belt objects, as well as theoretical models of gravitational interactions between the outer planets and a remnant planetesimal disk, provide strong evidence that the giant planet orbits migrated over large distances (up to many astronomical units (au)) after the solar nebula dissipated, with the orbits of Saturn, Uranus and Neptune expanding while that of Jupiter contracted[41–43]. Earth accretion simulations have explored the effects of this type of migration (considered by so-called Nice models), whose timing is uncertain[44–49]. It is possible that the giant planet orbits also underwent earlier migration before the nebula dissipated, driven by gravitational interactions between the planets and the gaseous nebula[50–52].

Although results vary across simulations that adopt different treatments and envisioned conditions, the models generally all predict a final phase of giant collisions. The Moon-forming Giant Impact, by an embryo often referred to as Theia, is thought to be the last of what were probably several giant impacts experienced by Earth during this phase[53], with a predicted occurrence time between 10 and 150 Myr (refs. 51,53).

Such theoretically predicted timescales must be reconciled with those determined from analysis of physical samples. The Hf/W ratio in the bulk silicate Earth (BSE) is about 20 times the average Solar System or chondritic value[54], so it should have a high relative abundance of ^{182}W from decay of ^{182}Hf if formed within the first 50 Myr of the Solar System (Box 2). Instead, the BSE's W isotopic composition is only slightly (about 200 ppm) more radiogenic than chondritic[46,47]. Therefore, much of the radiogenic ^{182}W formed by ^{182}Hf decay in the silicate Earth was removed by core formation, with the residue being diluted by protracted accretion over tens of millions of years[55]. The Moon also has a high Hf/W ratio[56], and the Moon and BSE have nearly identical W isotopic compositions. These isotopic compositions indicate that the Moon also formed late, >50 Myr after CAIs. Exactly how late is debated, as discussed below. Thus, there is broad agreement between isotopic constraints on the timing of Earth's formation and theoretical models.

Relating the W isotopic compositions of Earth and Moon to a more specific timescale for Earth's accretion is challenging. One approach is to adopt an idealized planetary growth model, such as an exponentially decreasing rate of accretion[1], to yield a mean age[55,57–59]. However, there might instead have been a long hiatus between a stage of early growth (10 Myr) and a late giant impact (>50 Myr)[55,60,61]. This hiatus is underconstrained, however, because the mean age of Earth calculated from all W isotope models is strongly dependent on the degree of equilibration of impactor metal with the silicate Earth, which is unknown, as discussed below[55,60,62–64].

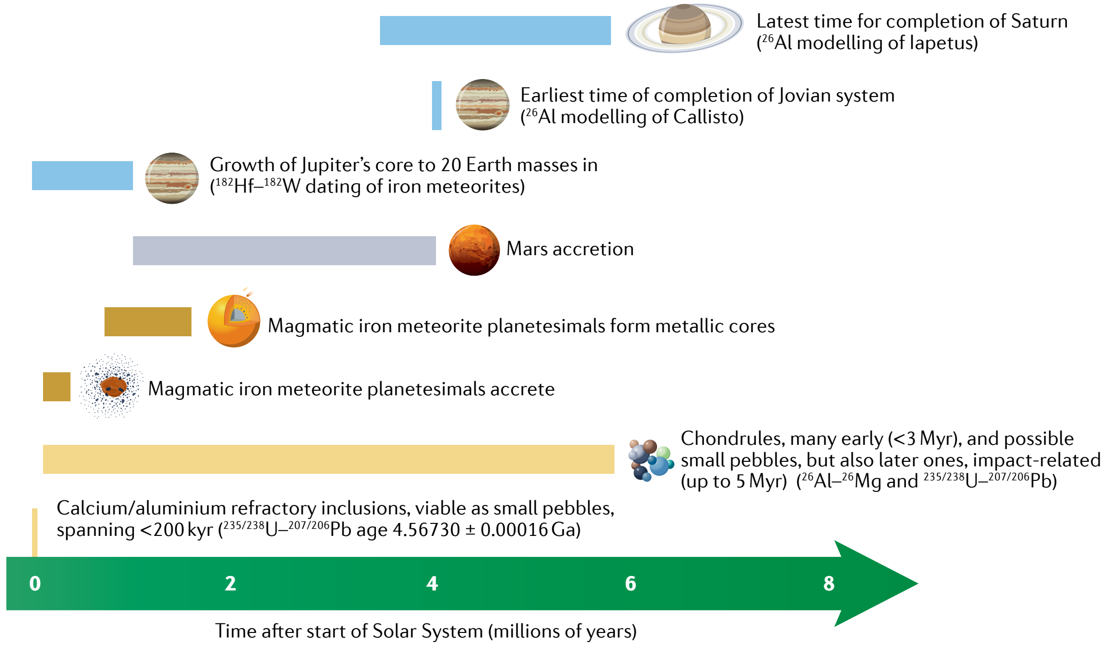


Fig. 2 | Chronology of the earliest Solar System, as determined from various isotopic systems. The oldest Solar System solids are the calcium–aluminium refractory inclusions (CAIs) forming at <200 kyr after the start of the Solar System. These were followed shortly after by the formation of chondrules that persisted for several million years. The timing of planetesimal accretion (~300–500 kyr) and differentiation (~800 kyr to 1.5 Myr) is inferred from the age of iron meteorites. Mars's accretion occurred early, probably within the first 4 million years. Jupiter's accretion occurred even faster in the presence of the gas-rich nebula, with its core growing to 20 Earth masses in the first million years. Data compiled from refs.[4,15,18,26,34–36,231]. Therefore, accretion of planetary objects, from dust to gas giants, was completed within less than 10 million years. In contrast to these early events, Earth's accretion was completed much later, around 70 to 120 Myr after the formation of CAIs. Diagram of planetesimal with metallic core adapted from ref.[232], Springer Nature Limited.

Although terrestrial accretion simulations are successful in accounting for Earth's mass and protracted accretion, the models tend to yield final planet orbits that are too eccentric compared with the actual terrestrial planets, a discrepancy revealed in the earliest *N*-body models[39,40] that still persists on average in modern works[65]. Essentially, the mutual embryo perturbations needed to yield crossing orbits and four final inner planets also tend to produce orbits that are too dynamically excited. Further, most simulations fail to reproduce Mars's small mass relative to the masses of Earth and Venus. Multiple potential solutions to this issue have been proposed[50,66–68], including histories in which either an early inward Grand Tack migration of Jupiter to about 1.5 au in the presence of the solar nebula[50], or a giant planet orbital instability during final assembly of the terrestrial planets[68], removed much of the mass in Mars's accretional zone. Distinguishing between these potential solutions remains challenging, in part owing to computational difficulty in resolving Mars's growth due to its small mass. It also remains unclear why the Solar System lacks planets interior to Mercury, whereas compact exoplanetary systems are common. Resolution of these issues requires more realistic models, and/or recognition that the terrestrial planets in the Solar System might reflect early events not typical of most planetary systems.

Although open questions remain, there now is a reasonably complete conceptual framework for how the initial circumstellar disk of gas and dust evolved to yield planetary building blocks and ultimately planets. This framework has been developed through a combination of theoretical modelling and increasingly high-fidelity analyses of physical samples from Earth, Moon, Mars and the asteroid belt. It is clear that Earth mainly formed late, after the nebula dissipated, and that its particular properties would have been affected by the stochastic nature of its final large accretionary impacts, as well as perhaps by the orbital migration of the outer giant planets. In the following sections, we explain how this protracted evolution led to Earth's composition.

Earth's composition

Earth's elemental and isotopic composition reflects where in the Solar System its various building blocks originated, as well as the processes involved in its construction. Meteorites, especially chondrites, provide fingerprints that help to identify the provenance of Earth's constituents (Box 1). These constraints, in combination with predictions of theoretical accretion models, can reveal how Earth acquired its chemical inventories and how they were modified. A challenge in unravelling this history is that current meteorite collections might not fully represent the range of compositions present when the inner planets accreted. Further, although samples from Earth, Mars and the Moon are available for detailed study, the bulk elemental and isotopic compositions of Venus and Mercury — reflecting the innermost disk accretionary conditions — remain unknown.

A refractory-enriched, volatile-depleted Earth

Although Earth was derived from material that formed in the circumstellar disk, its composition is distinctly non-solar. This compositional difference reflects a variety of factors, including melting, evaporation and condensation processes in the disk, loss of volatiles during and after accretion, and accretion of rocky components after the solar nebula dissipated. The most primitive (solar-like) chondrites are the CI carbonaceous chondrites (Box 1). The elemental concentrations in the BSE normalized to those in CI chondrites as a function of the 50% condensation temperature of each element in nebular conditions[69] (Fig. 3) provide evidence of the elements that have been lost to the core or to space, and which might have been enriched. Earth is relatively enriched in very refractory, lithophile elements (Hf, Ta and Ti); depleted in core-forming moderately siderophile refractory elements (Fe, Ni, Mo, W and Co); depleted in moderately volatile lithophile elements (K, Rb and Cs); more greatly and variably depleted in moderately volatile, moderately siderophile elements (Pb, Sb, Cd, Ag); extremely depleted in highly siderophile elements (Pt, Ir, Ru, Au, S and Te); and variably depleted in volatile, atmophile elements (C, N, H and noble gases).

Therefore, the BSE is enriched in the most refractory elements, as concentrated in CAIs (Box 1), and is depleted in highly and moderately volatile elements relative to CI chondrites, which themselves are depleted in the most volatile atmophile elements relative to solar. Temperatures in the early inner disk were initially high, so that disproportionately more refractory elements (condensed preferentially as CAIs for example) were ultimately incorporated into Earth through accretion of inner-disk embryos. These same conditions led to fewer volatiles being incorporated into inner embryos because of volatile loss when the gas disk dispersed[70,71].

At greater heliocentric distances, lower disk temperatures would have led to planetesimals and planetary embryos richer in volatiles. In general, material in the inner disk would have shorter collision times with Earth and thus have been accreted earlier than material sourced from more distant orbits[72–74]. Thus there has long been a view that the later stages of Earth accretion were more enriched in volatiles[75], which is supported by silver isotopes[76]. The BSE's high Pd/Ag ratio should have led to excess radiogenic ^{107}Ag from former decay of ^{107}Pd ($T_{½}$ = 6.5 Myr). Instead, the Ag composition is that expected from late-accreted, less volatile-depleted objects with low Pd/Ag ratio, such as are represented by chondrites.

Certain elements could have been lost from accreting embryos by collisional erosion of their outer portions[62,77–79]. For example, Earth's low total Mg/Fe ratio, relative to chondritic, is possibly explained by impact erosion of silicate after core formation[77], although loss of basaltic magma droplets and vapour to space during protoplanetary eruptions is an alternative explanation[80]. The experimentally determined volatilities of a number of elements within silicate liquids show that the patterns of depletion in Earth are more consistent with magma degassing and loss, rather than with the standard interpretation of incomplete condensation from the solar nebula[81]. Moderately volatile elements (Fig. 3) would have been particularly susceptible to such losses during early protoplanetary accretion when gravitational forces were smaller[79,82].

Earth's non-chondritic H, C and N ratios (Fig. 3) provide additional constraints on its accretion and losses of major highly volatile elements. With the advent of Moon-forming Giant Impact models (discussed below), it became common to presume that Earth's original water, carbon and nitrogen was lost and then redelivered as part of a subsequent chondritic veneer. The late delivery of water from comets or watery asteroidal objects[83] is dynamically plausible, as more distant material would impact Earth at later times[74]. However, late delivery of water is no longer considered likely, as theoretical work has since demonstrated that volatile loss was limited even during the Moon-forming impact[84]. This theoretical work, together with the non-chondritic H, C and N ratios (Fig. 3) and the particularly large depletion in nitrogen[85], provides evidence that these highly volatile budgets are not late additions but are largely residual to earlier accretion modified by differential loss, especially of nitrogen. Some have argued that the bulk of

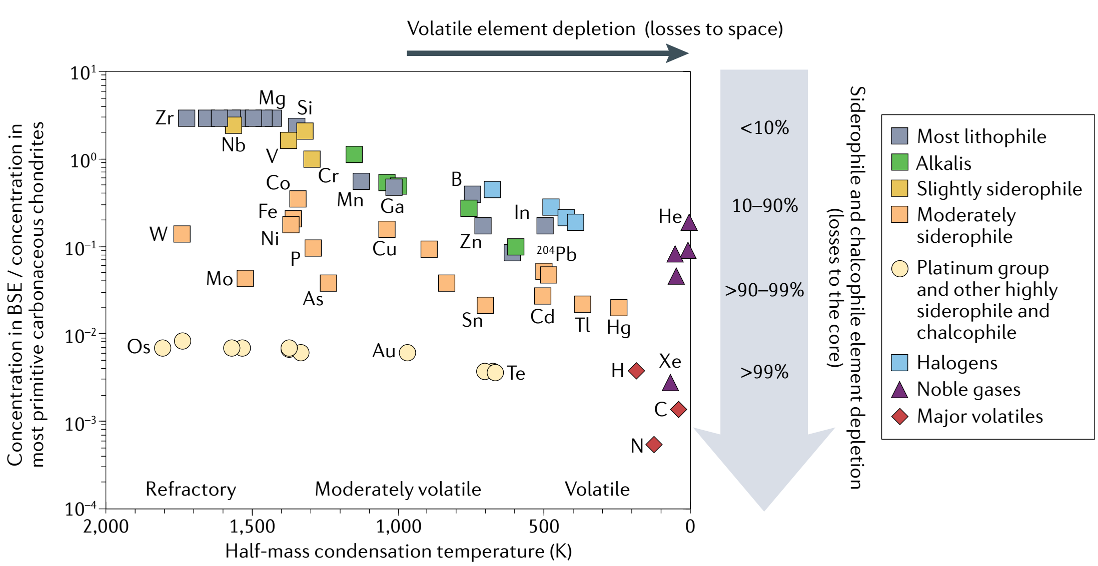


Fig. 3 | Concentrations of elements in bulk silicate Earth (BSE) relative to the most primitive carbonaceous chondritic abundances. Plotting normalized concentrations against revised half-mass condensation temperature[81] provides a picture of the fractionation behaviour of elements and groups of elements[69,81,85,233,234]. From left to right, the behaviour of elements goes from refractory to volatile. From top to bottom, the behaviour of elements goes from lithophile to siderophile and chalcophile. The silicate Earth is strongly depleted in moderately volatile and volatile elements but with striking variability in noble gases and major volatiles, including major depletion in Xe and N. Conversely, it is enriched in refractory elements, and the halogens seem far less depleted than previously thought[233]. The highly siderophile and chalcophile elements, which might be dominated by a late veneer or a flux from the core, define a pattern that is less depleted in moderately volatile elements than the lithophiles.

Glossary

Achondrite
Silicate-rich meteorites thought to be derived mainly from the outer portions (crust), rarely mantle, of other planetary objects, including Mars.

Astronomical units
(au). A distance of 150 million kilometres, the approximate distance from Earth to the Sun.

Atmophile
Elements preferentially concentrated in the atmosphere and hydrosphere, such as H, C, N and noble gases.

Bulk silicate Earth
(BSE). The integrated composition of the atmosphere, hydrosphere, crust and mantle, or simply Earth minus its core.

Chalcophile
Elements preferentially incorporated into sulfide and expected to be concentrated in the core, including S, Se and Te.

Evection
A gravitational resonance involving Earth, Moon and Sun that occurs when the period of precession of the lunar perigee matches Earth's orbital period around the Sun.

Highly siderophile
Elements preferentially (>99%) incorporated into metal and expected to be concentrated in the core, such as Ru, Pd, Re, Os, Pt and Ir.

Late veneer
The last material accreted to Earth post-Giant Impact, originally identified by the near-chondritic BSE abundances of the platinum group elements.

Lithophile
Elements preferentially (>50%) incorporated into silicate and expected to be concentrated in the mantle and crust, such as Si, Al, Ca, K, Mg.

Magmatic iron meteorites
Those iron meteorites with compositional trends and (in some cases) cooling textures indicative of core crystallization in a planetesimal or planetary embryo.

Moderately siderophile
Elements preferentially (50–99%) incorporated into metal and expected to be concentrated in the core, such as Ni, Co and Mo.

Moderately volatile
Elements that condense from a hot (solar) nebular gas at temperatures between 1,230 K and 640 K.

Moon-forming Giant Impact
The collision between the proto-Earth and another planet, often called Theia, that led to the formation of Earth's Moon.

Pebbles
Very early millimetre-to-decimetre-sized objects that formed in the earliest stages of the nebular disk by sticking together of dust grains and perhaps molten droplets.

Planetary embryo
Planetary objects of order 10^3 km in size formed by runaway growth from accreting planetesimals.

Planetesimals
Early 100-km-scale planetary objects that probably formed through the gravitational collapse of regions of dense concentrations of pebbles in the presence of the solar nebula.

Refractory elements
Elements that condense from a hot (solar) nebular gas at temperatures more than 1,400 K.

Solar nebula
The disk of gas and dust surrounding the newly forming Sun.

Volatile elements
Elements that condense from a hot (solar) nebular gas at temperatures less than 640 K.

Earth's H, C and N was accreted by the Moon-forming impactor[86,87], with Earth's high C/N ratio providing evidence of preferential N loss to the core[87] or to space[88]. D/H ratios for Earth and Moon are broadly consistent with carbonaceous chondritic sources consistent with the outer Solar System[89]. However, there is evidence that material of similar provenance to the enstatite chondrites, considered more proximal to Earth, could be the source of its water[90]. There also is evidence of an early H component in the deep Earth, acquired from the solar nebula[91].

Testing these hypotheses for the origins of Earth's volatiles is best achieved with the noble gases, which are powerful tracers of accretionary processes and sources. From their abundances and isotopic compositions, noble gases clearly represent a mixture of solar-composition gases derived from the nebula, as well chondritic and cometary components[85,92]. The ^{3}He/^{20}Ne ratio of Earth's mantle is greater than solar, providing evidence that nebular components could have been dissolved in a magma ocean in equilibrium with a hot vapour atmosphere[93]. Earth's core could also have added ^{3}He-enriched noble gases to the mantle over time[94–97]. The isotopic composition of Earth's interior and/or primordial Ne is consistent with a 75% solar and 25% chondritic mixture, whereas atmospheric Ne, although a vastly bigger reservoir, is only 20% solar[85]. This profoundly important heterogeneity[98,99] indicates a dominant, non-solar component acquired during main- and late-stage accretion.

Cometary contributions seem to have been important for the heavier noble gases. Earth's strikingly selective depletion in Xe (Fig. 3), and to a lesser extent Kr, could be explained by adsorption onto planetary ices or clathrates[100]. This ice adsorption has been confirmed with Xe abundance and isotopic data for cometary ices in 67P/Churyumov–Gerasimenko[101]. Xenon also has a sufficiently low first ionization potential that it could have been selectively removed and isotopically fractionated by extreme ultraviolet irradiation of the Archaean atmosphere[102], consistent with a change in atmospheric Xe isotopic composition over time[103].

Provenance of Earth's accreted components

The provenance of Earth's materials can be ascertained from the nucleosynthetic makeup of key elements (Box 1). The details of how the disk's nucleosynthetic architecture developed are becoming clearer with highly precise isotopic analyses[5,104,105] and elemental models[106]. That the provenance and nature of material accreting to Earth changed as the planet grew (Box 1) has been long appreciated in relation to chemical changes and bulk compositions[75]. However, more precise meaning has been brought to such arguments based on isotopic compositions[107]. Elements that are highly siderophile (Ru) are lost to the core throughout most of accretion (Fig. 3). Therefore, the isotopic composition of the residual element in the BSE reflects that of the last accreted material (Box 1). In contrast, elements that are lithophile (Ti) reflect Earth's full accretion history. Moderately siderophile elements (Ni) provide something in between. Isotopically speaking (excluding Si), Earth accreted from equal proportions of enstatite and ordinary chondrite-like material for the first half of its growth, with subsequent growth almost exclusively by material isotopically like enstatite chondrites[107]. Carbonaceous-chondrite-like materials are minor contributors (Box 1).

Despite the elegance of this solution, other interpretations of nucleosynthetic data have been proposed, and no meteorites, or combination thereof, exactly match Earth. For example, Mo isotopes have been used to indicate that carbonaceous chondrite material was delivered late to Earth, perhaps by Theia itself during the Giant Impact[86]. Further, it has been proposed that some of Earth's precursors were carbonaceous-chondrite-like materials that were severely fractionated isotopically and chemically by losses during early planetesimal formation, melting and evaporation[108]. Yet these proposals are hard to reconcile with multielement nucleosynthetic modelling that greatly limits involvement of carbonaceous chondrites (Box 1). Indeed, Ni isotope data provide evidence that materials accreted to Earth during the Giant Impact were enstatite-chondrite-like, leading to a proposal that Theia was a Mercury-like planet[109].

Even enstatite chondrites do not provide an exact match (Box 1). Recent Ru data[105] provide evidence (Box 1) of an unknown, more *s*-process-enriched endmember than carbonaceous, ordinary or enstatites in pre-veneer Earth formation. Because Ru is highly siderophile, these Ru data almost certainly constrain the source of Theia. Combined Mo and Nd data also show more generally that Earth's makeup requires a more extreme (but unknown) *s*-process-enriched component than is found in any chondrites[110].

Enstatite chondrites may be closest to Earth in their oxygen and some nucleosynthetic isotopic compositions (Box 1), but they are highly reduced and chemically unlike today's upper mantle. If Earth has such a composition[111], then the lower mantle must be compositionally distinct from the upper mantle. Such large-scale mantle heterogeneity is hard to reconcile with a long history of subducted slab penetration[112] but could relate to new evidence of Si-enriched domains in the lower mantle[113].

Enstatite chondrites also have Si that is isotopically fractionated relative to other objects[114,115], apparently limiting how much could be incorporated into Earth[116,117]. Some work, however, found that Si isotopic differences within EH chondrites arose from fractionation with metal in the solar nebula, and argued that a fractionated version might provide a viable component for Earth[118].

Earth is distinct from enstatite chondrites in terms of its bulk chemistry, and also is non-chondritic more broadly in Mg/Si and Al/Si ratios[119]. Although these could be affected by Si transfer to the core[120], it has been argued[121] that many of these differences represent chemical variations across the disk. Some think that this variation reflects disk-processing. It has been proposed[122] that the first planetesimals accreted from high-temperature condensate-rich (CAI-like) material with enhanced Al/Si and Mg/Si, and that these dominated the feedstock for Earth's accretion. Certain parts of this model are debated based on chondrite compositions[106] but it is consistent with Earth's enrichment in refractory elements (Fig. 3).

Various works[18,106,123–125] propose that chondrules could better represent the stock from which terrestrial planetesimals first accreted. Indeed, Mg isotope data provide a hint that Earth, Mars and Vesta are isotopically heavy, probably reflecting selective accretion from chondrulitic components that were melted in the circumstellar disk[124]. Advances have allowed Mg data of unparalleled precision to be obtained and combined with *N*-body simulation results to estimate melting and vaporization produced during accretionary collisions, indicating the resulting vapour–liquid isotopic fractionations expected above magma ponds[125]. For accretion histories involving a Grand Tack[50], predicted fractionations in Mg (and in similarly volatile Fe and Si) are consistent with those observed. Although it is uncertain that a Grand Tack occurred, the general mechanism is compelling, and thus Si and Fe mass fractionation could partly reflect early collisional heating[125] rather than solely core formation[126].

Heating and core formation

The ages of magmatic iron meteorites (Fig. 2; Boxes 1 and 2) indicate that melting and core formation occurred in early protoplanetary objects ≤10^2 km in size, which would have been fuelled by heating from decay of ^{26}Al ($T_{½}$ = 0.72 Myr) and ^{60}Fe ($T_{½}$ = 2.6 Myr). As accretion progressed, collisional heating became dominant, leading to melting and core formation in larger, later-formed bodies[127], and silicate magma oceans after particularly energetic impacts[128]. In a deep magma ocean, disproportionation of FeO into core-forming Fe metal and residual mantle Fe_2O_3 would have caused the silicate Earth to become more oxidized[129,130]. Changes in magma ocean oxidation state, temperature and pressure during Earth's growth would affect elemental partitioning into core-forming metal, with an evolution from reduced to more oxidized consistent with observed depletion in slightly siderophile elements such as Si, Nb and Cr (Fig. 2)[131]. However, Cr and V have been used to imply an evolution to less oxidizing conditions[132], consistent with isotope models[107].

As described above, Earth's main growth involved accretion of large planetary embryos, which would themselves have been differentiated with metallic cores. How impactor core material mixed within the BSE and affected its composition depends on the nature of each large collision, because collision geometry and energy affect the portion of impactor metal that fragments into small droplets within the BSE (via Rayleigh–Taylor instabilities) versus that which plummets as major inverted diapirs through the mantle as it descends towards the core[63,133,134]. The assumption of metal–silicate equilibrium during core formation[135] is only valid for impacting metallic material dispersed in small droplets in the magma ocean. If a substantial amount of impacting metal merged directly with Earth's core, it would change siderophile element concentrations and their isotopic compositions[62,63,134,136]. Certain features of the siderophile element geochemistry of Earth can best be explained in this fashion[64].

The late veneer

Dynamical models predict that after the last Giant Impact on Earth, a small portion of Earth's mass was delivered by much smaller impactors[51,53,137]. It has long been argued that a late veneer[138] of approximately 0.5% Earth masses was accreted after core formation, based primarily on the near-chondritic relative proportions of highly siderophile elements such as Ru, Pd, Re, Os, Pt and Ir in the BSE (Fig. 3), in contrast to the non-chondritic proportions expected if these elements had been incorporated into the core with their varied partitioning efficiencies. However, high pressures and temperatures associated with late core formation generally reduce differences in partitioning across the highly siderophile elements[139–143]. Further, with improved abundance data for the BSE it has become clear that some highly siderophile elements are not exactly chondritic in their relative proportions[144]. This non-chondritic highly siderophile element pattern has been explained in part by partial removal of some of the veneer in S-rich metallic liquids[145] associated with the final dregs of core formation[146,147].

Evidence suggests a proportionally much smaller late veneer on the Moon, which is challenging to understand. The proportion of late objects striking the Moon versus Earth is a function of encounter

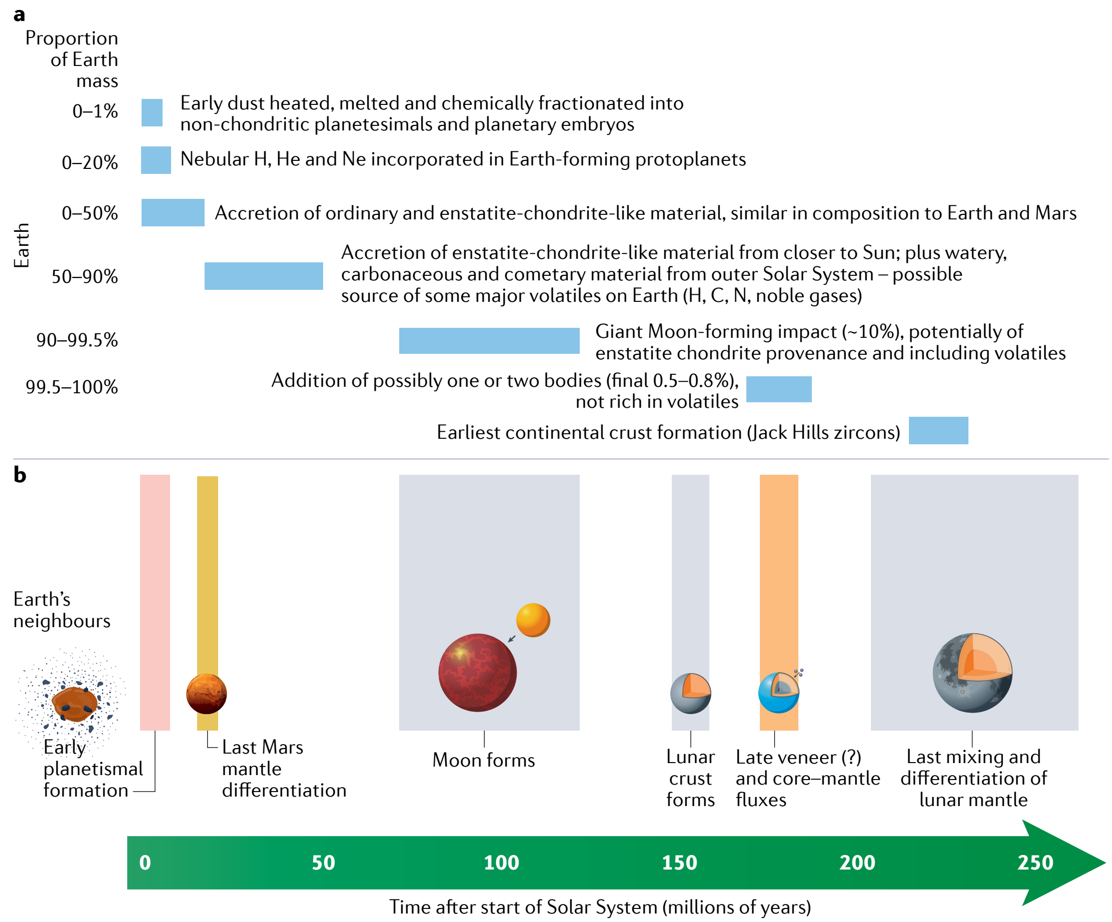


Fig. 4 | Best estimates for the chronology of Earth accretion. a, The stages of Earth accretion by percentage of Earth's mass. **b**, Comparison with stages of formation and differentiation of Mars and the Moon. Data shown are based on a variety of sources[15,18,26,34,56,60,76,104,107,147,154,155,163,165–171,229,231,235]. Earth almost certainly started accreting early and fast, like the other planetary objects. However, it continued for longer, culminating in the Moon-forming Giant Impact. Whether there was a substantial late veneer is now less clear with evidence that the fingerprint of highly siderophile elements in the silicate Earth could, in part, originate from a flux from the core.

velocity — for fast encounters, this proportion is dominated by the relative target cross-sections rather than by gravitational focusing by either object. The surface area of the Moon is about 7.4% that of Earth, whereas its mass is 1.2%. A pervasive late veneer would be expected to deliver a proportionally larger fraction of the Moon's mass than Earth's if their cross-sectional areas controlled the impact probability, whereas the fractional contributions would be more similar if gravity played the main role.

Neither of these is found: the Moon's late veneer (defined by highly siderophile abundances in basalts) is only 0.02–0.035% of its mass[148–150], a much smaller proportion than the 0.5–0.8% for Earth[144]. One idea is that Earth's late veneer was provided by the core of Theia, rather than by post-Giant Impact bombardment[151]. Others argue instead[152] that the Moon actually had a more massive late veneer that is no longer detectable, because some highly siderophile elements were sequestered into FeS (ref.[145]) and a lunar core. Alternatively, most or all of Earth's late veneer was delivered by only one or two large, lunar-sized bodies, rather than via a flux of smaller objects, so that it was statistically probable for the Moon to have avoided its proportionate share[153,154].

Late impactors of such size would introduce major perturbations to Earth's melting[154], as well as admixing materials that would affect isotopic compositions of the silicate Earth[134]. It has been argued that Earth's late veneer was added prior to the oldest surviving crustal zircons dated at 4.35 Ga (ref.[155]) and possibly as early as 4.45 Ga (ref.[154]) (Fig. 4). However, W and Pt isotope data for Isua and other ancient rocks[156–159] imply that the late veneer was not yet fully present, or not efficiently admixed, into the early Archaean mantle. That the veneer was not fully admixed by 4 Ga is surprising given a hotter, more rapidly convecting interior, not to mention melting expected from large collisions. However, modelling suggests that a major Hadean flux of impactors is less likely than previously thought[154]. If late additions were actually quite minor, maybe the veneer recorded in HSEs actually reflects gradual core contributions to the mantle over time.

In summary, Earth's chemical and isotopic compositions reflect a combination of the provenance of accreted disk components and their derivative planetary objects, which were diverse and changed over time. Representative materials are only partially represented by meteorite collections, but the latest isotopic research is leading to a mapping across the former disk. Modifications of protoplanetary materials led to compositions that are non-chondritic. The degree to which the mantle's chemical and isotopic compositions reflect these components further modified by equilibrium core formation, or disequilibrium resulting from merging of accreting planetary cores, is unclear. Finally, it has been difficult to find a straightforward explanation for the late veneer and how it was added to Earth's mantle over time, a subject we return to after discussing the origin of the Moon.

The origin of the Moon

Any viable model of Earth's assembly must account for the notably large Moon. Modern theories couple Moon formation to the last Giant Impact (or perhaps impacts) experienced by Earth as it completed its primary accretion. However, explaining the dynamical and detailed compositional features of the Earth–Moon system is challenging, and the nature of the Giant Impact best able to successfully account for these remains highly debated, as is its timing.

Late Moon formation

As discussed above, both Earth's and the Moon's W isotopic compositions imply that they completed their formation >50 Myr after CAIs. This was long after the accretion of Mars and the outer gas planets, which occurred within ~10 Myr after CAIs. These relative ages provide strong supporting evidence for the Giant Impact theory of Moon formation. After cosmogenic corrections, only a 20 to 30 ppm difference is resolvable between the W isotopic composition of the silicate Moon and the BSE[160,161]. This small W isotopic difference has been attributed to the preferential addition of the late veneer to Earth's mantle after Moon formation[161]. That the pre-veneer silicate Earth and Moon are inferred to have had essentially equal W isotopic compositions, together with the fact that there are no internal variations in W isotopic composition in the Moon, despite diverse source differences in Hf/W ratios, constrains the Moon's age to later than 70 Myr (ref.[161]). However, there is disagreement about this, and both earlier and much later ages have also been proposed.

An earlier Moon formation has been proposed[56] by authors who remodelled the Hf/W of the lunar mantle and concluded that it is significantly higher (30 to 50) than that of the BSE (25.5). They accounted for the small current difference in ^{182}W abundance between the BSE and bulk silicate Moon not in terms of a veneer[161] but purely as a result of lunar core formation while ^{182}Hf was extant, leading to excess radiogenic ^{182}W relative to the BSE, defining an age of <55 Myr. This reinterpretation has been questioned[162] based on the nature of the terrestrial late veneer[163], which in turn led to a compromise proposal that the BSE–Moon isotopic difference reflects a combination of a chondritic terrestrial late veneer and lunar core formation, leading to a revised age of <75 Myr (ref.[164]).

An early Giant Impact might appear more consistent with the I–Pu–Xe closure age of Earth's atmosphere of 30–60 Myr, inferred to date the Moon[165]. This closure age assumes, however, that the terrestrial atmosphere only reflects Earth's degassing history, whereas it actually integrates many different accreted components[85]. An age for the Moon of 60 ± 10 Myr based on Lu–Hf model ages of lunar zircons has also been proposed[166]. However, other work[162] points out that these Hf data exhibit a spread of values, with some initial compositions being less radiogenic than the Solar System initial, possibly reflecting cosmogenic effects. Therefore, there is no strong supporting evidence for an age earlier than 70 Myr[161]. However, there are arguments for a younger age.

An early Moon formation is hard to reconcile with ages of ferroan anorthosites, long thought to represent flotation cumulates from a lunar magma ocean resulting from the Giant Impact. Very precise ages close to 4.40 Ga (refs.[104,167]) have been obtained, consistent with U–Pb ages of lunar zircons of 4.417 ± 0.006 Ga (ref.[168]) (Fig. 4). These and similar results suggest a much later Moon formation[104,167], perhaps as late as 4.425 ± 0.025 Ga, or 115–165 Myr after CAIs[169]. A later age is also consistent with the Sr model age of 90 ± 20 Myr (ref.[56]), but this in turn is difficult to explain if the Moon is later than about 120 Myr.

Recognizing the conflicting arguments, it appears probable that the Moon formed between 70 and 120 Myr after the start of the Solar System, an age that is still consistent with W isotopic constraints of Earth's accretion. If the Giant Impact was in the earlier part of this range, the lunar magma ocean must have lasted longer than modelled[169], or lunar anorthosites might not have been formed in this way. There remains a preponderance of ages for lunar differentiation that are even younger, at around 4.34–4.37 Ga, or about 200 Myr after the start of the Solar System[170,171] (Fig. 4). These ages are younger than lunar zircon ages, but similar to the more reliable ages of Earth's oldest Hadean zircons[155] (Fig. 4). The 200 Myr lunar differentiation ages could be explained by a late major episode of melting and overturn in the Moon[170,171] and perhaps Earth[60,172]. Such late overturning could be associated with tidal effects[173] and/or a late major bolide striking Earth and contributing the late veneer[154].

Isotopic crisis for the Giant Impact model

A low-velocity, oblique collision by a Mars-sized protoplanet with the early Earth — a 'canonical' Giant Impact — can account for the Earth–Moon system angular momentum, as well as the Moon's mass and its bulk iron deficiency[174–176]. Most of Theia is absorbed by Earth, but a small portion achieves bound Earth-orbit, producing a disk that is derived predominantly from Theia, which later accretes into the Moon. However, the Moon and the BSE have essentially identical isotopic compositions for all non-volatile elements. If Theia was as isotopically different from Earth as meteorites from Mars (and most of those from parent bodies in the asteroid belt as well), one would expect a disk derived from Theia to yield a Moon with a measurably different isotopic composition than Earth, which is not seen, a quandary that has been referred to as an isotopic crisis for the Giant Impact theory[177].

The Moon and Earth do differ in their volatile element abundances, and likely in volatile isotopic compositions as well. The Moon's greater depletion in moderately volatile elements such as K and Rb could be explained by loss associated with a Giant Impact[178], and/or because volatiles initially in the prelunar disk were preferentially accreted by Earth rather than by the Moon[179,180]. The elements Cl, K, Zn, Rb and Ga are isotopically variable among lunar samples, with many being heavy relative to terrestrial[181–186]. Detailed comparisons between S, Cl and Zn with a common sample suite allow correction for effects associated with eruption and provide evidence that the Moon's moderately volatile elements are indeed slightly heavy isotopically[187], possibly caused by the Giant Impact or degassing from the magma ocean.

The Earth–Moon isotopic similarity in oxygen was apparent even in the Apollo data[188] and has received the most attention[189]. A possible explanation is that Theia was Earth-like in isotopes like O due to formation from a common region in the circumstellar disk. However, an additional explanation is needed to account for Earth–Moon Si and W isotopic similarities, which reflect other, more individually varying, accretional processes.

The isotopic composition of silicon is unusually heavy in the BSE relative to chondrites and most other meteorites[114]. Nebular fractionations are likely to generate considerable Si isotope heterogeneity in the innermost Solar System[118]. Enstatite chondrites, in particular, are isotopically light[114,116,117]. The heavy BSE signature is distinctive and could, at least in part, reflect isotopic fractionation during high-pressure core formation[114,126,190–192]. Yet the identical isotopic signature is found for the Moon[115], within which high pressures would not have occurred. An isotopically heavy Si feature has also been found in angrites, which would not have experienced high pressures, but in this case it could reflect extreme volatile depletion of the parent body[193].

The W isotopic composition reflects the time-integrated Hf/W ratio, which also is a function of core formation conditions as well as timing[55]. Indeed, the W isotopic compositions of basaltic achondrites, formed by melting of parent body mantles, are all different. Yet the initial Earth and Moon are inferred to have been identical, with the now resolvable very small Earth–Moon difference reflecting the late veneer and/or a small lunar radiogenic component[56,150,162,164]. Accounting for equal Earth–Moon W isotopic compositions is particularly challenging for Giant Impact models. Not only would the silicate mantles of Theia and the proto-Earth be expected to have differing W compositions, but during the impact the metal from Theia's core becomes highly disrupted. Some probably plummets toward Earth's core without equilibrating with the mantle[62–64]. However, a substantial amount remains dispersed in the mantle, and some is lofted into the prelunar disk. Even a small amount of this W-rich material, relatively unradiogenic in terms of ^{182}W, would markedly alter the isotopic composition of Earth or its resultant moon.

Giant Impact models

A variety of explanations have been offered to reconcile Moon formation via a Giant Impact with Earth–Moon isotopic similarities (ref.[194] and references therein; Table 1). A first class of models involve modifications to a canonical impact by requiring an isotopically Earth-like Theia[107], or efficient mixing and compositional equilibration of disk and Earth mantle material after the impact but before the Moon accreted[189,195], or that Earth's mantle at the time of the impact was molten causing a proportionally larger fraction of Earth material to be ejected into the prelunar disk[196]. For each idea, open issues remain. An isotopically Earth-like Theia for many elements is plausible[107,197–199], but this does not predict the Si similarity, and it appears wholly improbable as an explanation for W (refs.[161,200]). Equilibration is appealing because it could account for a wide range of Earth–Moon isotopic similarities, including W (ref.[201]), but whether it would operate sufficiently before the Moon forms is uncertain. The cited effects of a molten mantle[196] could depend sensitively on details of the numerical method[202] and have not been replicated in other works[203].

A second class of models considers impacts with much higher angular momentum, with larger impactors and/or a fast-spinning-Earth before the giant impact. It was previously thought that the Earth–Moon system's angular momentum would have remained nearly constant since the Moon formed, changing by only a few per cent due to later impacts and tides raised by the Sun on Earth[153,175]. A discovery in 2012 was that dynamical interactions with the Sun through a well-known resonance (evection) could have greatly slowed the spin of the early Earth, transferring angular momentum from the Earth–Moon pair to Earth's heliocentric orbit[204]. The Earth–Moon angular momentum just after the Moon-forming impact could then have been up to 2 to 3 times the current value. Various high-angular momentum and/or high-energy impacts have been considered, including those that can directly produce a disk and planet with nearly equal isotopic compositions in O and other lithophile elements[204,205], as well as intermediate cases that require some equilibration[195]. Such impacts produce a highly vaporized planet whose angular momentum is so large that the planet expands until its equatorial velocity matches the local orbital velocity, so that there is then no difference in velocity between the planet's outer edge and the inner edge of the circumplanetary disk, a structure that has been termed a synestia[195,206]. Whether and how the needed angular momentum removal would occur remains an actively debated issue[207–210].

Other models consider hit-and-run impacts, in which much of Theia escapes after a higher-velocity giant impact, which ultimately leads to a more Earth-like disk via the initial impact[211] or after a secondary collision[212]. Such cases might still require modest equilibration or a relatively isotopically Earth-like Theia.

Perhaps the most distinct idea holds that the Moon is the product of tens of collisions involving sub-Mars impactors[213]. Each impact creates a moonlet that tidally migrates outward, merging with other moonlets produced by earlier impacts. The Moon is then built up through many such events, with the final Earth–Moon compositions approaching that of the mean planetesimal neighbourhood. However, the merger efficiency between consecutive moonlets is low[214], with most moonlets lost to collision with Earth, so that creation of a lunar-sized Moon via this mechanism appears improbable.

Both the timing and nature of the Moon-forming Giant Impact thus remain uncertain. Various different models have been proposed to account for the close isotopic similarity between the BSE and the Moon, which invoke very different impact conditions (Table 1). Distinguishing among such varied ideas on Moon formation is crucial to understanding the conditions of Earth's final assembly.

Table 1 | Comparing Moon-forming Giant Impact scenarios

Example Giant Impact scenarios	$M_{Theia}/M_{\oplus}$ [a]	Velocity (v_{esc}) [b]
Canonical impact[174–176] and equilibration[189]	0.13 to 0.2	1 to 1.2
Canonical impact[174–176] and Earth-like Theia[107,198]	0.13 to 0.2	1 to 1.2
Canonical impact[174–176] and Earth magma ocean[196,203]	0.13 to 0.2	1 to 1.2
Hit-and-run impact[211,212]	0.2 to 0.3	1.2 to 1.4
High-angular-momentum impact with fast-spinning proto-Earth[195,204]	0.03 to 0.1	1.5 to 3
High-angular-momentum impact between two half-Earths[195,205]	0.4 to 0.5	1 to 1.5
High-angular-momentum impact and equilibration[206]	0.03 to 0.5	1 to 3
Multiple impacts[213]	0.01 to 0.1	1 to 3

Predicted Theia mass and impact speed across various Giant Impact scenarios. A wide range of impactor masses and Giant Impact energies are proposed by different theoretical and numerical models, with varied implications for the post-Giant Impact initial states of Earth and Moon. [a]Mass of Theia (M_{Theia}) relative to present-day Earth's mass ($M_{\oplus}$). [b]Velocity in units of mutual escape velocity (v_{esc}).

Mantle isotopic heterogeneities

An exciting development in understanding Earth's accretion has been the discovery of mantle isotopic heterogeneities that can only have formed in the first ~100 Myr of Earth history. These findings and their implications are summarized here.

The convecting terrestrial mantle carries isotopic heterogeneities only up to about 2 Ga in age generated by decay of long-lived radionuclides (Box 2). It had long been assumed therefore that primordial heterogeneities resulting from decay of short-lived (<100 Myr) nuclides would have been completely homogenized by mantle mixing. Instead, it was demonstrated in 1999 that there was an isotopic difference between the Xe in well gases and that from mid-ocean-ridge basalt (MORB) sources[215]. Later work[216] showed that this isotopic difference extended to Xe differences between MORB and ocean island basalts. The parent ^{129}I half-life of 16 Myr means that these heterogeneities were established within the first 100 Myr of Solar System history – roughly the most commonly accepted time of the Moon-forming Giant Impact.

There is also a difference in ^{142}Nd abundance between early Archaean sedimentary rocks and those from Earth's mantle[217–221], thought to reflect decay of formerly live ^{146}Sm (half-life 68 ± 7 Myr)[222]. It has been argued[172] that this Nd isotopic difference might also reflect early heterogeneity residual from the Moon-forming Giant Impact, assuming that the Giant Impact did not occur until 4.45 to 4.35 Ga.

Excesses of ^{182}W have also been found in Precambrian rocks, which taken in isolation could be explained by incomplete mixing after the Giant Impact subsequently fully homogenized in today's mantle following billions of years of convection. However, deficits in ^{182}W abundance have also been found in modern-day plume basalts and kimberlites[223], interpreted as residual heterogeneities from accretion and core formation[224,225], or even core-derived components[226,227] that have persisted until today. Those W isotope variations that are well defined exhibit relationships with other isotopic parameters including a negative correlation with ^{3}He/^{4}He (ref.[227]). That is, the ^{182}W deficiency is associated with domains that have been thought of as less depleted in primordial noble gases and thus more likely to reflect the earliest Earth.

The association between ^{3}He excesses and ^{182}W deficits raises expectations of other correlations with siderophile and chalcophile elements[228] and possibly radiogenic Xe anomalies, mandating an intensity of profoundly important new research over the coming years. It has been proposed that mantle Xe isotopic anomalies are, as with more and less radiogenic W isotope anomalies, related to episodes of core formation in which iodine becomes siderophile under high pressures[229]. This siderophile behaviour of I and Xe would be most readily associated with the inevitable high-pressure core formation following the Giant Impact.

Any relationship with the Giant Impact is problematic, however, since the ^{182}Hf half-life requires that, depending on the Hf/W ratio, such heterogeneity was generated within the first 50 Myr. However, this early age of the Giant Impact would be hard to reconcile with latest estimates of Moon's age (Fig. 4). Furthermore, it is surprising that Earth's mantle was not fully mixed and homogenized by the Moon-forming impact. If it were not fully mixed, this would provide a powerful constraint on Giant Impact models[230].

An alternative is that the heterogeneity was introduced by a large, differentiated projectile or two, contributing the late veneer[134]. However, the mantle today is heterogeneous in ^{182}W/^{184}W, but not in ^{142}Nd/^{144}Nd, from decay of formerly live ^{146}Sm. The variability in ^{142}Nd/^{144}Nd seen in Archaean rocks was homogenized by the Proterozoic Eon, making it unlikely that tungsten could have avoided such mixing, so that slow mixing of a late veneer is an unlikely explanation for the range in ^{182}W/^{184}W seen in modern ocean island basalts.

A third interpretation is that a plume component derived from the core carries both primordial ^{3}He and unradiogenic W (refs.[94–97,226,227]), a slow flux of which would decouple W from Nd. It would also generate, at least partially, Earth's apparent late veneer of highly siderophile and/or chalcophile elements. However, a plume component would imply that the present-day Ru isotopic composition of the silicate Earth reflects that of the outer core, which makes the differing early and late compositions harder to understand (Box 1).

Therefore, after decades of research developed within the paradigm of a well-mixed mantle, following the addition of a late veneer, ancient mantle W isotopic anomalies are providing exciting evidence that is causing a major rethink. The evidence of plume-derived ^{3}He-enriched components with unradiogenic W implies that a long-term flux of deep mantle and/or core components could have led to some if not all of the 'late veneer'. This deep mantle and/or core flux into the mantle over geological time would explain the difference between the silicate Earth and Moon in W isotopes but also in the abundances of highly siderophile elements. It would also be consistent with the differing longevity of early heterogeneity between W and Nd.

Summary and future directions

The post-Apollo era has witnessed an epoch of explosive growth in understanding Earth's formation, fuelled by three main factors: the acquisition of lunar samples and increasing success in the collection of meteorites worldwide; ground-breaking developments in mass spectrometry, providing extraordinary precision in characterizing isotopic compositions; and a substantial increase in computational power. It is now possible to model planet assembly in a near-experimental sense, using large-scale computer simulations to produce a statistically significant number of predicted outcomes that can be directly compared against the growing wealth of physical constraints, with discrepancies revealed through such comparisons then feeding back into the creation of ever more realistic theoretical and numerical models.

A major open question involving how small grains that initially orbited in the Sun's preplanetary disk were able to accumulate into large building blocks of planets – a first essential step in Earth's origin – appears to have been solved theoretically, through increased understanding of the collective behaviour of self-gravitating particles as they interact with the primordial gaseous nebula. However, reconciling such predictions with meteoritic evidence remains incomplete. Earth's bulk chemical composition seems to better reflect processed materials such as chondrules, rather than bulk chondrites. Yet, most (but not all) chondrules appear late (~3 Ma), possibly formed by flash melting from shocks in the disk, so that the role of chondrules as protoliths for early accretion remains unclear. More evidence is needed on the nature of very early chondrule- or pebble-forming events and the way in which the nebular silicate liquid and gas separated. Continued observations of protoplanetary disks will also be essential, allowing for the study of large numbers of young disks at varied stages of accretion, as well as comparison with statistics of exoplanetary systems.

Earth's assembly was protracted, lasting from 70 to 120 Ma, and it is known that accretionary processes would have been affected by dynamical interactions with the earlier-formed outer giant planets, whose orbits evolved greatly during or soon after terrestrial accretion. There is broad agreement that the Moon originated as a result of a giant collision (or perhaps more than one) with Earth at the end of its primary growth, reflective of a phase of impacts between protoplanets as the inner Solar System settled into a dynamically stable final state. However, there exists a remarkable range of Giant Impact models, and distinguishing among them will require both modelling advances and new lunar data. Theoretical models that link impact conditions to observable chemical properties of the Moon will be important, although such predictions might not prove unique. Two other questions regarding the early Moon – whether it formed in a fully or only partially molten state, and how its orbital inclination was acquired and maintained – appear more challenging to address, and could ultimately better distinguish between theoretical concepts. New lunar samples originating from the deep crust and/or mantle that better reveal the Moon's bulk composition, as well as new geochemical and/or geophysical constraints on the depth of the initial lunar magma ocean, are key new data needed for progress. Prospects for acquiring such data are promising, given recommended lunar missions for NASA by the Decadal Survey in Planetary Science and Astrobiology[231], as well as international plans for lunar exploration.

Throughout terrestrial planet assembly, settling of denser metallic phases to planetary cores left compositional signatures in the remnant planetary mantles, which today provide constraints on the timing and nature of planetary growth. However, both require a more sophisticated understanding of the complexities of mixing and equilibration of silicate and metal solids and liquids during the range of extreme pressures and temperatures associated with planetary collisions, requiring advances in simulation techniques and experimental constraints.

Remarkably – despite Earth's molten initial state and tectonically active interior – some preserved heterogeneities in its interior date back to its formation at >4.5 Ga. These need to be explained to identify whether the core is a substantial source of noble gases and responsible for mantle He, Ne, Xe and W isotopic heterogeneity, and to find to what extent the late veneer, defined by highly siderophile element abundances, could instead be a product of core–mantle fluxes over geological time. Further, relative to chondrites, nitrogen is the most depleted element in the silicate Earth. It is unclear how and why nitrogen was lost, but better experimental and theoretical constraints from theoretical models of losses to the core or space are needed.

Finally, the multielement approach to understanding Earth's nucleosynthetic makeup has provided a powerful framing of accretion. However, given that meteorites do not reproduce Earth's isotopic composition exactly (and that meteorites might only represent a surviving subset of original nebular compositions), the provenance of Earth's *s*-process component remains unclear (Box 1). There also is a major mismatch between the nucleosynthetic and chemical compositions of Earth. Sample return missions and/or in situ isotopic analyses from Venus and Mercury – which together comprise nearly half of the total mass in the inner Solar System, but whose detailed bulk and isotopic compositions remain a mystery – would yield transformative advances in understanding.

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Additional information

Correspondence should be addressed to Alex N. Halliday.

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