***Making Planets on Earth: How Experimental Petrology Is Essential to Planetary Exploration***

**Primary Author:**

Kayla Iacovino1,\*

**Co-authors:**

Nicole G. Lunning2,Gordon M. Moore1, Kathleen Vander Kaaden1, Francis M. McCubbin2

**Co-signers:**

Jack P. Touran1, Jeremy Boyce2, Kevin Righter2

\*Phone: 480-265-6372, email: kayla.iacovino@nasa.gov

1Jacobs, NASA Johnson Space Center, Houston, TX 77058, USA

2NASA Johnson Space Center, Houston, TX 77058, USA

**1.0 Introduction**

Our knowledge of the chemical evolution of the solar system relies upon our ability to interpret a limited set of observations, including geochemical data from returned samples and meteorites, spectral and other remotely sensed data from spacecraft and telescopes, and geochemical information measured *in situ* via landers and rovers. Our ability to obtain these data is constantly improving in quality and quantity, but the understanding of the universe derived from these data is only as good as our interpretive tools. Experimental petrology is the most powerful way in which we define fundamental geochemical principles, e.g., thermodynamic equations of state, volatile solubilities and partitioning, partitioning of elements between gaseous, fluid, silicate and/or metal phases, and pressure-temperature-redox-composition controls on phase composition and stability. These laboratory derived principles are critical to precisely interpreting and converting hard-won geochemical data sets into constrained models of the formation and evolution of our solar system.

 Experimental petrology is a laboratory technique where the conditions within planetary bodies – from cores to atmospheres – are reproduced and measured in a controlled environment. In this way, it is possible to investigate the precise pressure (*P*), temperature (*T*), redox (*f*O2), and compositional (*X*) conditions under which natural rocks formed. There are two fundamental services that experimental petrology provides. First, the intensive parameters of specific rock formations (P, T, *f*O2, X) may be determined by “reverse engineering” the conditions that produce certain rock characteristics (e.g., mineral assemblage, major and minor element concentration and partitioning, etc.). This can be useful for understanding the evolution of a rock or suite of rocks and how they relate to the planetary body as a whole. Second, experiments can be performed using a broader strategy, in which swaths of geochemical space (e.g., *P*-*T*, *T*-time, *X*-time) can be mapped empirically. This, in turn, provides two scientific products: 1) fundamental relationships between geochemical variables can be established for specific systems (e.g. meteorite systems, planetary surfaces, etc); and, 2). over time, all of these studies can be used as database input for thermodynamic models capable of predicting rock formation conditions (e.g., MELTS). Both of these products are critical to developing and constraining our understanding of the fundamental processes that built and modified our solar system over the last 4.5 billion years.

**1.1 Providing a foundation for interpretation**

Experimental petrology is critical to planetary science as a foundational tool that facilitates connections between disparate observational datasets and allows for holistic interpretation of those datasets in terms of geologic processes. All of the ways in which we obtain data require a foundation for data interpretation and analysis. Remotely sensed geochemical data of planetary surfaces requires an understanding of atmospheric chemistry and low-pressure mineral phase equilibria. Linking observations of hand samples (meteorites and eventual sample return) to the geologic processes that formed them requires knowledge of petrologic evolution of a multitude of geochemical systems and phase assemblages (e.g., volatile, silicate, metal). Interpretation of the chemical makeup of planetary crusts requires constraints on planetary-scale differentiation processes, including element partitioning between silicate mantles and metallic cores. Petrologic experimentation provides a ground-truth empirical and thermodynamic foundation on which to build interpretations of these astromaterials.

 In the following sections, we detail the various experimental methodologies employed to investigate each major geologic regime on planetary bodies. These sections are organized by pressure regime, planetary atmospheres through surfaces, crusts, mantles, and deep within planetary cores. Examples where experimental petrology has formed the basis of major advancements in our understanding of the universe are given, exemplifying the ongoing need for petrologic experimentation to be a continuously funded line of planetary science research in the decade to come. In each section, we highlight gaps in our knowledge and outline how experimental work should be expanded or created to meet new research goals.

**2.1. Space weathering**

 Rocks on the surfaces of bodies with thin to no atmospheres experience geochemical alteration that is attributed to interaction with solar wind and micrometeorite impacts, processes commonly referred to as space weathering. These airless bodies include a broad range of planetary objects at a variety of distances from the Sun (e.g., the Moon, asteroids, and some outer planet rocky moons). The last decade of research has revealed the effects of space weathering may vary based on distance from the sun. It has been well established that space weathering modifies different mineral assemblages and chemical compositions differently, but we still do not understand specific alteration processes for many common surface materials1. Ongoing and upcoming spacecraft missions to a wide variety of airless bodies, including OSIRIS-REx, Hayabusa2, MMX, DART, Psyche, and Lucy, will certainly reveal more about space weathering processes across the solar system. New experiments examining and constraining the space weathering conditions relevant to these bodies will be required to understand the anticipated observations from these important missions.

Two types of experiments have been used to explore space weathering on a variety of materials: irradiation with hydrogen ions to replicate solar wind modification2; and pulsed laser experiments to replicate micrometeorite bombardment effects3. These experiments have replicated the production of nanophase-Fe0 and agglutinates recognized as space weathering effects in lunar regolith samples, and nanophase-Fe0 recognized in particles from the near-Earth asteroid Itokawa4,5. Recent space weathering experiments on carbonaceous chondrites indicate their visible-near-IR spectra and mineralogy may change in different ways than occurs for lunar surface samples6. The main belt asteroid Vesta does not have visible-near-IR spectra with reddening associated with lunar-style space weathering and solar-wind-bearing howardites thought to be surface samples from Vesta do not appear to contain nanophase-Fe0, indicating that Vesta experienced space weathering differently than the Moon even though they have similar minerals present on their surfaces. Currently, we do not know how space weathering affects many of the materials we observe remotely on the surfaces of airless bodies. Space weathering experiments need to be conducted on additional types of surface analogs to explore how recognized space weathering processes affect known surfaces. This testing may reveal additional space weathering processes, or how the effects of solar wind and micrometeorite impacts combine in different ways as heliocentric distance varies.

**2.2 Planetary atmospheres and atmosphere-fluid-rock interactions**

Planetary atmospheres can range from relatively intert (e.g., contemporary Mars) to relatively reactive (e.g., Venus), having radically differing consequences for the chemical compositions of the uppermost layers of rocks on the surface of other planets. The compositions of atmosphere and fluids have important implications for habitability conditions on Europa, Titan, and in the past on Mars. As we learn more about current and past atmospheres of these bodies over the next decade, it will be possible to run targeted laboratory experiments to expand our understanding of the geochemical processes at work. Understanding how planetary atmospheres and fluids interact with each other and solid surface materials is crucial for determining the surface evolution of planetary bodies. In addition, atmosphere-surface fluid, rock-atmosphere, ice-atmosphere, and fluid-rock interactions can be geochemically preserved in surface rock deposits from which complementary experimental work can be used to elucidate past atmosphere and/or surface fluid compositions.

Aqueous alteration processes played an important role on some carbonaceous chondrite parent bodies and modified the organic building blocks for life found in some meteorites7,8. We anticipate learning more about these organic building blocks from the samples that will be returned by OSIRIS-REx and Hayabusa2. In addition, improvements in our telescopic observation capabilities (e.g., the James Webb Space Telescope) may strengthen connections between carbonaceous chondrite meteorite groups and specific asteroids, which would allow for estimation of parent body sizes and potential pressure constraints for early solar system aqueous alteration processes. The potential discoveries above will warrant experiments that incorporate these new constraints to understand the conditions needed to preserve specific suites of organic components and to advance our understanding of processes related to water in the early solar system.

Recreating and understanding atmospheric environments aids in our interpretation of alteration of rocks on the surfaces of planetary bodies. Crucially, this uppermost rocky layer is the *only* portion of planetary bodies routinely measured via remote sensing and even rover and lander operations. As such, understanding the reactions taking place in planetary atmospheres has a direct bearing on the interpretation of the principal geochemical observations we have of other bodies.

 Planetary atmospheres ranging from vacuum levels up to several bars can be recreated with an array of various instrumentation. The most common method is the reproduction of atmospheres at 1 bar of pressure (ambient pressure on the surface of the Earth) in non-pressurized gas-mixing furnaces. Such experiments have been commonplace for decades and, because thermodynamic properties of gases are so well known at 1 bar, form the empirical backbone of datasets extrapolated (or interpolated) to higher or lower pressures. Higher pressure atmospheres up to several bars can be simulated in closed reaction vessels, in which reagents are combined and sealed in a small, constant volume chamber which is then heated. Pressure is achieved internally, caused by the increase in volume in the reagents during heating. Experiments from 1 to several bars of pressure have been used to advance our understanding of gas-fluid-rock interactions on planetesimals and other planetary bodies, but these pressures are higher than the surface pressure on Mars, Mercury, and the exteriors of planetesimals. Some of the reactions occurring on these bodies are likely to be very pressure sensitive, and so to fully understand these worlds, we need to conduct experiments at the relevant pressures, below that of the ambient Earth. Vacuum furnaces and evacuated, sealed silica-glass tubes have been used to understand nebular condensation and degassing processes, including stable isotope fractionation, but these apparatuses have primarily been used to study the residual melt and infer the products that were lost via mass balance. Recent advances in the experimental community have enabled the development of sub-bar experimental systems designed to capture the residual melt, condensed solids that form from the degassed vapor, and any uncondensed gas. This “sub-bar furnace” consists of a vacuum system connected to an unpressurized tube furnace. Glass vacuum lines are run directly from the sample into a gas chromatograph, such that measurements of outgassing can be measured *in situ* at high temperatures or captured within a cold trap for subsequent analysis. Experiments from this system would enable our community to explore completely new questions about the gas and/or fluid compositions that exsolve from rocks or lavas on planetary bodies that have atmospheric pressures below 1 bar, such as Mercury, Mars, and differentiated asteroids.

**2.3 Planetary crusts and small bodies**

The igneous crusts of rocky bodies dramatically vary in their geochemical compositions across the solar system. The extremes of this crustal compositional range are arguably the Moon’s feldspathic highlands, Earth’s granitoid continental shields, and Mercury’s MgO-rich and FeO-poor boninitic crust, as recently revealed by the MESSENGER mission9–11. As a counterbalance to those crustal extremes, crusts with basaltic compositions are common; Venus, Mars, and the asteroid 4 Vesta have mostly basaltic composition crusts, and the lunar Mare and terrestrial oceanic crust are also composed of basalt12. Basaltic meteorites and human-collected lunar samples in our collections have varying minor, trace element, and isotopic compositions that suggest dramatic differences in the geochemical source regions that formed basalt on separate rocky bodies across the solar system. This implies an array of unique processes driving planetary geochemical evolution throughout the solar system. In recent years, possible lunar and Martian granites have been identified in meteorites and/or via remote sensing13,14. The formation processes required to form granites, even in small quantities, on these planetary bodies remain enigmatic. Over the next decade experimental petrology is one of our primary tools available to determine what these rock suites mean for crustal formation on their respective bodies.

Guided by meteorite and surface compositions derived from remotely sensed datasets, experimental studies of planetary igneous crusts have largely been undertaken in either (a) 1-bar gas-mixing furnaces (0.1 MPa) or (b) piston cylinder apparatuses at pressures of 500-3,000 MPa. Respectively, these experimental approaches simulate surface-to-very-near-surface depths and upper-mantle depths. These studies firmly established that petrologic reservoirs are unique to certain planetary bodies, and that redox conditions vary between different planetary bodies. These are both important factors that contribute to distinct crustal compositions for different rocky solar system bodies. Recent experimental approaches have sought novel ways to study the magmatic processes at intermediate depths between the surface and upper mantle, particularly to investigate pressure sensitive components such as volatiles that could influence atmospheric composition of their planetary body or fluid compositions that may alter shallow crustal environments15. These include experiments in 1-bar furnaces that capture chemical species that are lost from the experimental magmas due to volatilization15. Experiments at pressures which correspond to intermediate depths between the surface and upper mantle could substantially advance our understanding of volatile retention/degassing, as well as potential corresponding magmatic signatures. Experiments at these intermediate pressures (0.1­–400 MPa) are accessible with cold seal experimental apparatuses and are widely used by the terrestrial experimental petrology community but have only been used in few studies of extraterrestrial petrology.

The pressure range for the crustal depths of approximately Earth-sized planets (0.1–1,000 MPa) encompasses the full range of pressures within interiors of small solar system bodies, such as planetesimals and their contemporary descendants in the asteroid belt. For example, an 8-km-diameter rocky asteroid has a central interior pressure of 0.1 MPa (1 bar) and 500-km-diameter rocky asteroid (Vesta-sized) has a central interior pressure of 100 MPa. Experimental petrology is particularly important for understanding planetesimals because our ability to observe most of these bodies is limited to telescopic observations, exploration by only a handful of spacecraft (although there will be more in the next decade), and incomplete sampling of these bodies as in-falling meteorites. For example, we have iron meteorites that represent the cores of at least 50 separate differentiated planetesimals but we do not have crustal or mantle samples from that many bodies in our meteorite collections. Previously, experimental studies have been pivotal for understanding the formation of basalts and more Si-rich magmas by partial melting of chondritic starting materials. Experimental studies continue to help us to explain the processes that planetesimals underwent in the early solar system and to identify and determine the properties of the components missing from our sample collections.

In the coming decade, the Psyche Mission will explore the metallic asteroid 16 Psyche that is thought to be the core of a differentiated planetesimal. The Psyche mission is anticipated to reveal the composition of the outer core of its planetesimal, and some recent astronomical observations suggest core-mantle boundary regions may be exposed on the surface of this metallic asteroid. These anticipated observations will provide humanity’s first look at the interior structures of differentiated planetesimals, which will be a jumping off point for experiments aimed at understanding the differentiation processes that formed Psyche and how planetesimal core solidification proceeded (e.g., some models include inward core solidification)16. The model for differentiation and core solidification for Psyche will then be compared with additional meteorites and new experiments to understand the diversity (or lack thereof) in differentiation and core solidification processes in the early solar system.

**2.4 Planetary mantles**

Petrologic experiments provide geochemical information to complement what we know about Earth’s interior structure (mantle and core) from geophysical measurements. Currently, our knowledge of the interior structures of most planetary bodies is very limited. We can glean some information about the interiors of planetary bodies from their sizes, densities, and rotational parameters, but without in situ geophysical measurements we have very little direct knowledge about planetary interior structures. The anticipated results from the Insight mission may provide new constraints on the interior structure of Mars, which will provide the basis for more firmly establishing an interior structure of Mars.

With experiments, we can use the most primitive (mantle-derived) basalt compositions to reverse engineer the type of mantle material that would have undergone partial melting to form that specific type of basalt. This approach, naturally, will be improved when new primitive basaltic rocks are found as meteorites, brought back as returned samples, or geochemically analyzed in situ by rovers or landers. Currently, reverse engineering mantle compositions from Martian meteorite compositions indicates there are two geochemically distinct mantle reservoirs, but there remain many open questions about what these different mantle reservoirs indicate about the martian interior17.

Experiments on planetary mantles are typically carried out in solid-media pressure apparatuses capable of generating the relevant high pressures (piston cylinders and multi-anvils). These greatly differ from gas or fluid pressurized apparatuses used at low pressures in that samples are enclosed within a multi-component (typically ceramic) assembly and pressurized by applying force in one or multiple directions using hydraulic presses. The extreme higher pressures and temperatures found in planetary mantle systems make these experiments extremely difficult to perform, with high failure rates and steep thermal gradients across samples. Despite these challenges, high-pressure experimentation remains one of the foremost ways in which we constrain the deepest planetary processes, from differentiation and core formation to the interpretation of seismic and geodetic data. The measurement of the most fundamental thermodynamic parameters at these extreme pressures form the backbone of more complex modeling of these systems. It may be possible to generate these data at relevant pressures at less extreme temperatures, by altering the chemical composition of the chosen starting materials. Even if these compositions are less reflective of natural ones, it would enable data generation in regions of P-T space never before accessible, and the properties being measured may be compositionally indifferent. Over the next decade, experimental petrologists may take cues from the material science community, which investigates fundamental material properties that are less sensitive to the bulk composition of a material.

**2.5 Planetary cores: Metal-Silicate systems**

In bodies large enough to undergo differentiation into a metallic core and silicate mantle and crust, understanding the nature and rate of chemical and isotopic equilibration of elements between metal and silicate is critical for interpreting chemical, chronologic, and genetic data in planetary science. Metal-silicate partitioning for several key elements (typically siderophile elements such as Hf, W, Mo, Ru, Cr, Sn, Pt, etc.) has been studied extensively using experimental techniques18–20. To constrain a large *P*-*T* space, experiments of this type may utilize a variety of the experimental techniques described in this paper, from 1 bar (0.1 MPa) up to thousands of MPa.

 Of particular interest in ongoing and future metal-silicate studies is in constraining the partitioning behavior and equilibration timescales related to *isotopic* equilibrium of siderophile elements between silicate and metal. Currently isotopic experimental data are incomplete or absent. As a result, modeling the extent of equilibrium during accretion modeling of inner solar system bodies varies from full isotopic equilibrium to very limited equilibrium. For example, isotopic heterogeneities observed in the terrestrial mantle and derivative samples (basalts, komatiites) could be attributed to the addition of late veneer material to the Earth after core formation, or they could reflect a lack of isotopic equilibrium over the timeframe of mantle evolution21. Isotopic data in planetary materials have been measured for a wide range of siderophile elements including Mo, W, Ru, Cr, Sn, Pt, Zn, Cu, V, Os, S, and Si22–33. Without the corresponding thermodynamic information (i.e., partition coefficients and equilibration timescales) however, interpreting these data is difficult. In order to fully leverage our existing and continually growing datasets, we suggest that future experimental studies of metal-silicate systems should target the isotopic partioning and equilibration of these critical elements.

**3.0 Concluding remarks**

The success of missions like BepiColombo, Mars 2020, MMX, OSIRIS-REx, Hayabusa2, and Psyche over the next decade will provide us will unprecedented geochemical and petrologic data of non-gaseous planetary bodies across our solar system. In order to not only prepare for the interpretations of these complex datasets, but to enhance the scientific return of this data from these various missions, a wide range of experimental petrology studies are warranted; a select few were detailed in this paper. Given that petrologic experimentation provides a ground-truth empirical and thermodynamic foundation on which to build interpretations of these various datasets to truly understand the thermochemical and petrologic evolution of non-gaseous bodies across our solar system, it is imperative that this remain an active and funded line of research in the coming decade.

**References**

1. Pieters, C. M. & Noble, S. K. Space weathering on airless bodies. *J Geophys Res Planets* **121**, 1865–1884 (2016).

2. Dukes, C. A., Baragiola, R. A. & McFadden, L. A. Surface modification of olivine by H + and He + bombardment. *J Geophys Res Planets* **104**, 1865–1872 (1999).

3. Loeffler, M. J., Dukes, C. A., Christoffersen, R. & Baragiola, R. A. Space weathering of silicates simulated by successive laser irradiation: In situ reflectance measurements of Fo 90 , Fo 99+ , and SiO 2. *Meteorit Planet Sci* **51**, 261–275 (2016).

4. Noguchi, T. *et al.* Incipient space weathering observed on the surface of Itokawa dust particles. *Sci New York N Y* **333**, 1121–5 (2011).

5. Thompson, M. S., Christoffersen, R., Zega, T. J. & Keller, L. P. Microchemical and structural evidence for space weathering in soils from asteroid Itokawa. *Earth Planets Space* **66**, 89 (2014).

6. Thompson, M. S., Loeffler, M. J., Morris, R. V., Keller, L. P. & Christoffersen, R. Spectral and chemical effects of simulated space weathering of the Murchison CM2 carbonaceous chondrite. *Icarus* **319**, 499–511 (2019).

7. Lipschutz, M. E. & Schultz, L. *Meteorites in Encyclopedia of the Solar System (Second Edition), Chapter 13*. (2007).

8. Brearley & A.J. The Action of Water. *Meteorites and the early solar system II* **943**, (2006).

9. Korotev, R. L., Jolliff, B. L., Zeigler, R. A., Gillis, J. J. & Haskin, L. A. Feldspathic lunar meteorites and their implications for compositional remote sensing of the lunar surface and the composition of the lunar crust. *Geochim Cosmochim Ac* **67**, 4895–4923 (2003).

10. Kaaden, K. E. V. & McCubbin, F. M. The origin of boninites on Mercury: An experimental study of the northern volcanic plains lavas. *Geochim Cosmochim Ac* **173**, 246–263 (2016).

11. Taylor, S. R. & McLennan, S. M. The geochemical evolution of the continental crust. *Rev Geophys* **33**, 241 (1995).

12. Condie & K.C. Comparative Planetary Evolution in Earth as an Evolving Planetary System (Third Edition), Chapter 10. in 317–367 (n.d.).

13. Seddio, S. M., Jolliff, B. L., Korotev, R. L. & Zeigler, R. A. Petrology and geochemistry of lunar granite 12032,366-19 and implications for lunar granite petrogenesis. *Am Mineral* **98**, 1697–1713 (2013).

14. Wray, J. J. *et al.* Prolonged magmatic activity on Mars inferred from the detection of felsic rocks. *Nat Geosci* **6**, 1013–1017 (2013).

15. Ustunisik, G., Nekvasil, H. & Lindsley, D. Differential degassing of H2O, Cl, F, and S: Potential effects on lunar apatite. *Am Mineral* **96**, 1650–1653 (2011).

16. Scheinberg, A., Elkins-Tanton, L. T., Schubert, G. & Bercovici, D. Core solidification and dynamo evolution in a mantle-stripped planetesimal. *J Geophys Res Planets* **121**, 2–20 (2016).

17. Barnes, J. J. *et al.* Multiple early-formed water reservoirs in the interior of Mars. *Nat Geosci* **13**, 260–264 (2020).

18. Righter, K. & Shearer, C. K. Magmatic fractionation of Hf and W: constraints on the timing of core formation and differentiation in the Moon and Mars. *Geochim Cosmochim Ac* **67**, 2497–2507 (2003).

19. Touboul, M., Walker, D., Ash, R. D., Puchtel, I. S. & Walker, R. J. Simultaneous experimental determination of metal-silicate partitioning of W, Mo, Ru, Pt, and Pd using natural abundances, elevated P-T and isotopic tracers. *Lunar and Planetary Science Conference* **42**, 1727 (2011).

20. Wade, J., Wood, B. J. & Tuff, J. Metal–silicate partitioning of Mo and W at high pressures and temperatures: Evidence for late accretion of sulphur to the Earth. *Geochim Cosmochim Ac* **85**, 58–74 (2012).

21. Trønnes, R. G. *et al.* Core formation, mantle differentiation and core-mantle interaction within Earth and the terrestrial planets. *Tectonophysics* **760**, 165–198 (2019).

22. Herzog, G. F., Moynier, F., Albarède, F. & Berezhnoy, A. A. Isotopic and elemental abundances of copper and zinc in lunar samples, Zagami, Pele’s hairs, and a terrestrial basalt. *Geochim Cosmochim Ac* **73**, 5884–5904 (2009).

23. Savage, P. S., Armytage, R. M. G., Georg, R. B. & Halliday, A. N. High temperature silicon isotope geochemistry. *Lithos* **190–191**, 500–519 (2014).

24. Creech, J. B. *et al.* Late accretion history of the terrestrial planets inferred from platinum stable isotopes. *Geochem Perspectives Lett* 94–104 (2016) doi:10.7185/geochemlet.1710.

25. Goderis, S., Brandon, A. D., Mayer, B. & Humayun, M. Osmium isotopic homogeneity in the CK carbonaceous chondrites. *Geochim Cosmochim Ac* **216**, 8–27 (2017).

26. Kruijer, T. S. & Kleine, T. Tungsten isotopes and the origin of the Moon. *Earth Planet Sc Lett* **475**, 15–24 (2017).

27. Labidi, J., Farquhar, J., Alexander, C. M. O., Eldridge, D. L. & Oduro, H. Mass independent sulfur isotope signatures in CMs: Implications for sulfur chemistry in the early solar system. *Geochim Cosmochim Ac* **196**, 326–350 (2017).

28. Bermingham, K. R., Worsham, E. A. & Walker, R. J. New insights into Mo and Ru isotope variation in the nebula and terrestrial planet accretionary genetics. *Earth Planet Sc Lett* **487**, 221–229 (2018).

29. Sossi, P. A., Moynier, F. & Zuilen, K. van. Volatile loss following cooling and accretion of the Moon revealed by chromium isotopes. *P Natl Acad Sci Usa* **115**, 10920–10925 (2018).

30. Sossi, P. A., Nebel, O., O’Neill, H. St. C. & Moynier, F. Zinc isotope composition of the Earth and its behaviour during planetary accretion. *Chem Geol* **477**, 73–84 (2018).

31. Wang, X., Amet, Q., Fitoussi, C. & Bourdon, B. Tin isotope fractionation during magmatic processes and the isotope composition of the bulk silicate Earth. *Geochim Cosmochim Ac* **228**, 320–335 (2018).

32. Budde, G., Burkhardt, C. & Kleine, T. Molybdenum isotopic evidence for the late accretion of outer Solar System material to Earth. *Nat Astronomy* **3**, 736–741 (2019).

33. Nielsen, S. G. *et al.* Nucleosynthetic vanadium isotope heterogeneity of the early solar system recorded in chondritic meteorites. *Earth Planet Sc Lett* **505**, 131–140 (2019).