REVIEW SUMMARY

LUNAR GEOLOGY

The lunar dynamo

Benjamin P. Weiss[1*] and **Sonia M. Tikoo**[2,3]

BACKGROUND: It is unknown whether the Moon has a fully differentiated and melted structure with a metallic core or retains a partially primordial, unmelted interior. The differentiation history of the Moon is manifested by its record of past magnetism (paleomagnetism). Although the Moon today does not have a global magnetic field, the discovery of remanent magnetization in lunar rocks and in the lunar crust demonstrated that there was a substantial lunar surface field billions of years ago. However, the origin, intensity, and lifetime of this field have been uncertain. As a result, it has been unclear whether this magnetization was produced by a dynamo in the Moon's advecting metallic core or by fields generated externally to the Moon. Establishing whether the Moon formed a core dynamo would have major implications for understanding its interior structure, thermal history, and mechanism of formation, as well as for our understanding of the physics of planetary magnetic field generation.

ADVANCES: A new generation of laboratory magnetic studies of lunar rocks and spacecraft measurements of lunar crustal magnetic fields have produced major advances in our understanding of the evolution of ancient magnetic fields on the Moon. It has now been established that a dynamo magnetic field likely existed on the Moon from at least 4.5 billion to 3.56 billion years ago, with an intensity similar to that at the surface of Earth today. The field then declined by at least an order of magnitude by 3.3 billion years ago. The early epoch of high field intensities may require an exceptionally energetic power source such as mechanical stirring from mantle precession. The extended history of the lunar dynamo appears to demand long-lived power sources such as mantle precession and core crystallization.

OUTLOOK: Measurements of the intensity of the ancient lunar dynamo have shown that it was surprisingly intense and long-lived. The next phase of lunar magnetic exploration will be to obtain more accurate measurements of field paleointensities and to determine when the dynamo initiated and finally disappeared. This will be coupled with the continued development of magnetohydrodynamic models for characterizing mechanical and other unusual dynamo mechanisms and further investigations into the thermal, structural, and geodynamical history of the lunar core and mantle. The eventual availability of absolutely oriented samples and in situ spacecraft measurements of bedrock should enable the first measurements of the paleo-orientation of lunar magnetic fields. Such directional data could determine the lunar field's geometry and reversal frequency, as well as constrain ancient local and global-scale tectonic events. ■

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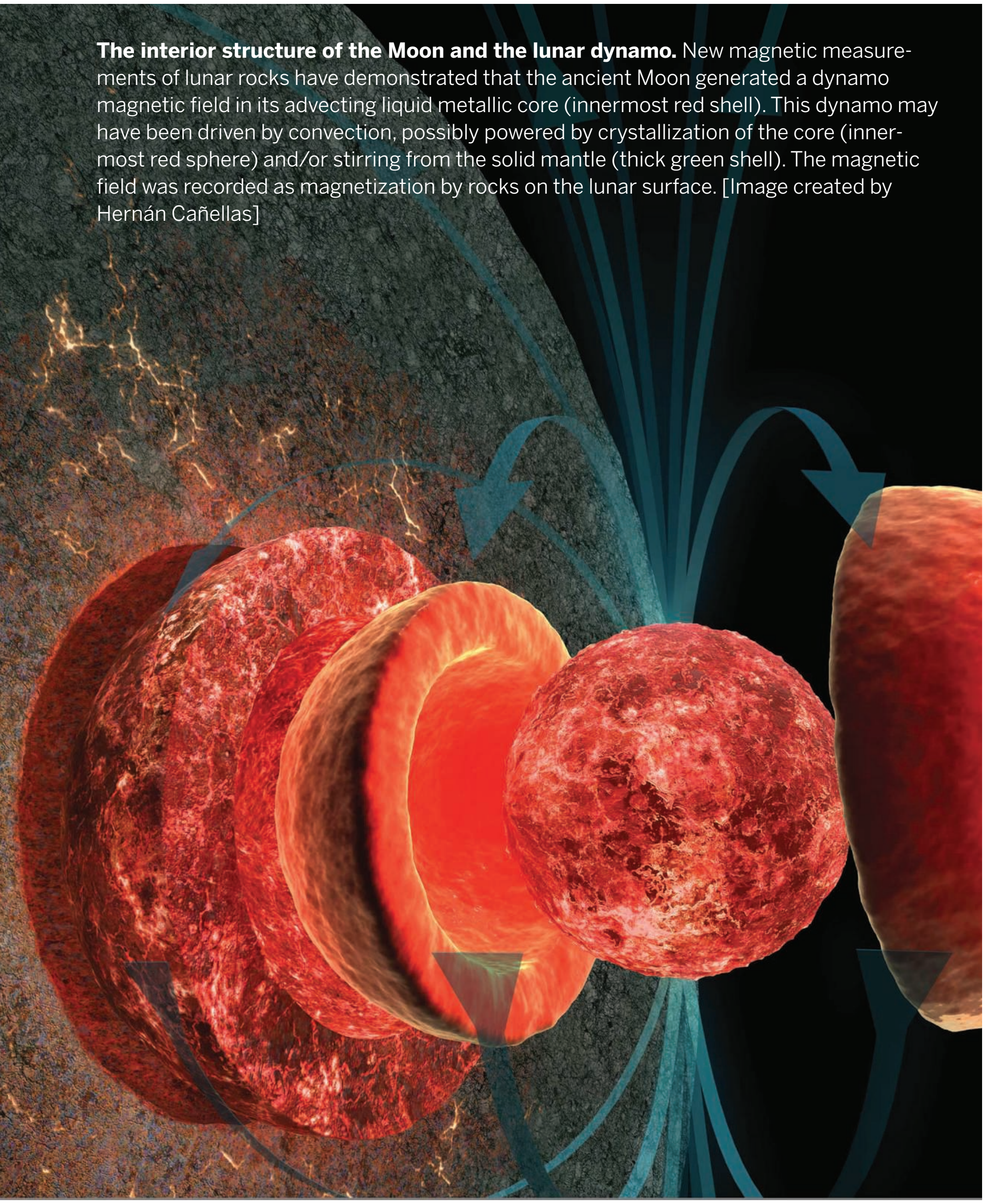

The interior structure of the Moon and the lunar dynamo. New magnetic measurements of lunar rocks have demonstrated that the ancient Moon generated a dynamo magnetic field in its advecting liquid metallic core (innermost red shell). This dynamo may have been driven by convection, possibly powered by crystallization of the core (innermost red sphere) and/or stirring from the solid mantle (thick green shell). The magnetic field was recorded as magnetization by rocks on the lunar surface. [Image created by Hernán Cañellas]

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The inductive generation of magnetic fields in fluid planetary interiors is known as the dynamo process. Although the Moon today has no global magnetic field, it has been known since the Apollo era that the lunar rocks and crust are magnetized. Until recently, it was unclear whether this magnetization was the product of a core dynamo or fields generated externally to the Moon. New laboratory and spacecraft measurements strongly indicate that much of this magnetization is the product of an ancient core dynamo. The dynamo field persisted from at least 4.25 to 3.56 billion years ago (Ga), with an intensity reaching that of the present Earth. The field then declined by at least an order of magnitude by ~3.3 Ga. The mechanisms for sustaining such an intense and long-lived dynamo are uncertain but may include mechanical stirring by the mantle and core crystallization.

One of the most important events in planetary history is early global differentiation, the large-scale melting and sequestration of compositionally distinct materials into a radially layered structure with a central metallic core and silicate mantle. The terrestrial planets inevitably differentiated during the first ~10 million to 100 million years (My) of their history due to their large sizes (radii > 2400 km), which led to the release of large quantities of gravitational energy during their accretion (*1*). By comparison, the gravitational energy of formation of even the largest asteroids (radii < 500 km) is only sufficient to have globally heated these bodies by up to few tens of degrees Celsius.

The fate of the Moon, which has an intermediate radius (1737 km), is less certain: Formation by gradual accretion of small planetesimals would not heat the body above the solidus (melting temperature) unless most energy was retained during accretion. In contrast, formation by the impact of a Mars- or larger-sized planetesimal on the early Earth would likely produce a Moon at or above the solidus (*2*). The latter giant impact scenario might naturally lead to the formation of an early molten core and dynamo magnetic field (*3*).

The Moon today does not have a global magnetic field, but the Apollo-era discoveries of remanent magnetization in the lunar crust and returned samples demonstrated that there were substantial lunar surface fields billions of years ago (*4*). However, it has long been unclear whether this magnetization was produced by a core dynamo and/or by magnetic fields generated externally to the Moon. Establishing whether the Moon formed a metallic core and dynamo would have major implications for the Moon's origin, igneous and thermal history, and large-scale interior structure. Furthermore, the Moon's small size makes it a unique natural laboratory for studying the physics of dynamo generation in a parameter regime not represented by other known planetary bodies. In particular, the apparent persistence and intensity of lunar magnetic fields are a powerful and potentially illuminating challenge to dynamo theory.

Over the past decade, a burst of experimental and theoretical advances has greatly improved our understanding of the lunar paleomagnetic record, global geophysical properties, and dynamo theory. Here, we review how these coupled discoveries have confirmed the existence of an ancient lunar dynamo, defined its temporal evolution and lifetime, and stimulated a new view of planetary magnetic field generation.

Scientific understanding of the Moon by the end of the Apollo era

Before the Apollo missions, there were two disparate views of the nature of the lunar interior. Urey (*5*) conceived of the Moon as largely a primordial, undifferentiated relic from the early solar system, whereas Kuiper (*6*) and others conceived of the Moon as a differentiated body with endogenic volcanism and a metallic core. Early magnetometry measurements by the Luna 1 and 2 spacecraft appeared consistent with Urey's view, finding no evidence of an intrinsic lunar field (*7*); we now know that the lunar dipole moment is ≤ 8 × 10^{-9} that of the Earth [using recent Lunar Prospector spacecraft measurements and the sequential model of (*8*)]. Furthermore, early measurements of the Moon's gravity and physical libration (i.e., three-dimensional temporally varying rotation) observed that its normalized polar moment of inertia appeared indistinguishable from that of a uniform sphere

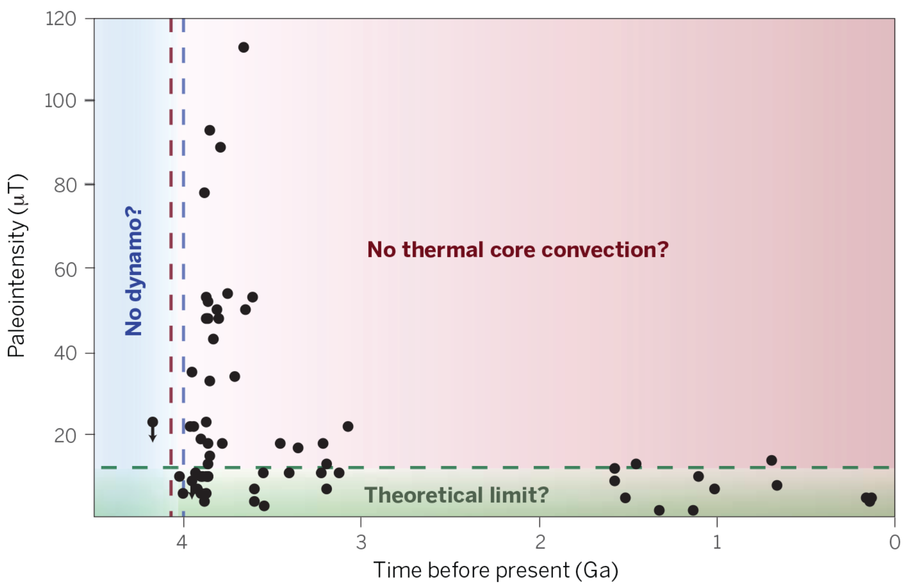

Fig. 1. Apollo-era paleointensity measurements of the lunar magnetic field and their perceived difficulties. Each point represents the inferred field intensity for a single lunar rock. Points with arrows represent upper limits. Both the intensity values and their ages are uncertain. The apparent rapid rise in paleointensities between 4.0 and 3.9 Ga has been taken as evidence that the origin of the lunar dynamo, and perhaps of the core itself, were delayed by 500 My (*3*) (blue), apparently at odds with a giant impact origin of the Moon. Energy flux scaling (Eq. 2) predicts maximum paleointensities of only less than ~12 μT (green), apparently well below many measured values. Early thermal evolution models (*25*, *27*) predicted that a core dynamo powered by purely thermal convection is unlikely to have persisted beyond ~4.1 Ga, before virtually all measured paleointensities (red). The weakest surface fields today are <0.2 nT (*29*). Paleointensities are obtained from the IRM normalization method as compiled by (*4*, *17*).

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(for which $C/MR^2 = 0.4$, where C is the polar moment of inertia and M and R are the lunar mass and radius). However, the subsequent discovery that the mare plains are basaltic by Surveyor 5 (9) and that the lunar highlands are anorthositic by Apollo 11 (10) confirmed that, beginning with an early post-accretional large-scale magma ocean, the Moon repeatedly experienced igneous processing over >2 billion years. Furthermore, improved spacecraft tracking in the mid-1970s established that $C/MR^2 \sim 0.392 \pm 0.003$ (11), consistent with (although not requiring) the existence of a small metallic core (1 to 4% of the lunar mass) (12, 13). Finally, the discovery of natural remanent magnetization (NRM) in the lunar crust and Apollo samples from spacecraft and laboratory measurements hinted at the possibility of an ancient but now extinct lunar dynamo.

Paleomagnetic measurement methodology

NRM is a vector quantity that reflects the macroscopic, semipermanent alignment of electron spins within ferromagnetic minerals. The major ferromagnetic minerals in unbrecciated lunar rocks are the metals kamacite (α-$Fe_{1-x}Ni_x$ for $x < \sim 0.05$) and martensite (α_2-$Fe_{1-x}Ni_x$ for $\sim 0.05 < x < \sim 0.25$) (4, 14, 15). Igneous rocks become magnetized when they cool in the presence of an ambient magnetic field below their Curie temperature (780°C for kamacite), a form of NRM known as thermoremanent magnetization (TRM). For weak planetary fields, the TRM intensity, M, is given by

$$M = \chi_{\mathrm{TRM}} B_{\mathrm{paleo}} \qquad (1)$$

where B_{paleo} is the ancient magnetizing field and χ_{TRM} is the TRM susceptibility, a sample-dependent constant that quantifies the abundance and magnetization efficiency of the ferromagnetic grains. Other forms of NRM, such as crystallization remanent magnetization (CRM) (acquired when ferromagnetic minerals crystallize) or shock remanent magnetization (SRM) (acquired when rocks experience transient high pressures), are presently more difficult to relate quantitatively to ancient field intensities. NRM in typical planetary materials can persist for longer than the age of the solar system, well after any ancient field has decayed away.

Because essentially all returned lunar rocks are regolith samples, their original orientations are largely unknown (16). Therefore, laboratory measurements of Apollo samples have usually sought to obtain the paleointensity, and not paleodirection, of past lunar fields. To obtain B_{paleo}, the first step is to measure the quantity on the left side of Eq. 1, M_{NRM} [after removing any partial magnetization overprints using alternating field (AF) or thermal demagnetization]. Then, because χ_{TRM} is difficult to measure, the sample is remagnetized in a known laboratory field, B_{lab}, and the resulting magnetization, M_{lab}, is measured. If χ_{TRM} is unchanged by this process, then the two resulting equations can be divided to obtain $M_{\mathrm{NRM}}/M_{\mathrm{lab}} = B_{\mathrm{paleo}}/B_{\mathrm{lab}}$ and then solved for B_{paleo}. In principle, the most accurate paleointensity estimates for igneous rocks could be obtained if M_{lab} were produced as a TRM, because this would resemble the form of the NRM. Doing so is the basis of Thellier-Thellier paleointensity experiments. However, heating-induced chemical alteration, combined with magnetostatic interactions between ferromagnetic grains, can change χ_{TRM} in a way that is difficult to quantify, leading to experimental failure (4). Such problems have frustrated Thellier-Thellier experiments of lunar rocks for decades, such that essentially none have been able to meet the standards of success typical for Earth rocks (17, 18). As a result, most paleointensities have been inferred using nonheating methods to produce M_{lab}. In particular, either a saturating field [to produce an isothermal remanent magnetization (IRM)] or weak field superposed on a strong AF [to produce an anhysteretic remanent magnetization (ARM)] is applied to the samples as a proxy for TRM. Using these methods, B_{paleo} can usually be estimated with a factor of ~5 uncertainty (2 SD) (tables S2 to S4).

Apollo-era measurements

By the end of the Apollo era (the early 1980s), analyses of hundreds of samples seemed to indicate the existence of ancient magnetic fields with intensities ranging from 0.1 to 120 μT over the period <200 million years ago (Ma) up to 4.0 billion years ago (Ga) (4) (Fig. 1). This huge range of intensities far exceeds the expected surface variations for a constant, selenocentric (i.e., Moon-centered) dipole moment, suggesting that the Moon's magnetic field may have experienced dramatic temporal variations (19).

Despite the wealth of data, there was no consensus on the origin or meaning of lunar magnetization. One reason is that many samples were complex breccias that experienced multiple shock and brecciation events (Fig. 2). Not only is it uncertain when such rocks became magnetized, but it is often not clear that the NRM being measured is actually a TRM (17, 18). For some samples, it was apparent that the NRM was contaminated during sample handling by the astronauts and/or after return to Earth (14, 20, 21). Furthermore, even apparently unshocked materials (i.e., those that never experienced peak pressures exceeding the 5-GPa threshold at which petrographic effects become evident), which include most mare basalts, yielded uncertain results because their constituent ferromagnetic minerals are usually in the form of multidomain crystals that have poor magnetic recording properties. As a result of these problems, both the age of the NRMs and the associated paleointensities were uncertain for the majority of analyzed samples.

Attempting to account for the nominal paleointensity record from a theoretical perspective is also very challenging. Perhaps most surprising is that the inferred paleointensities for many lunar samples are up to 1 to 2 orders of magnitude greater than the maximum surface fields predicted by dynamo scaling laws. For example, analysis of a diversity of numerical simulations of Earth-like dynamos powered by convection suggests that the mean dipole magnetic field intensity at the Moon's surface may be estimated as

$$B \approx d\left(\frac{R_c}{R}\right)^3 (2\mu_0 c f \rho)^{1/2}\left(\frac{\alpha G}{3C_P}\right)^{1/3}(4\pi R_c^2 q_a)^{1/3} \qquad (2)$$

where d is the ratio of the dipolar field to the total field at the core-mantle boundary, R_c is the core radius, R is the planetary radius, μ_0 is the magnetic permeability of free space, f is the ratio of ohmic dissipation to total dissipation, ρ is the core density, α is the volumetric coefficient of thermal expansion for the core, G is the gravitational constant, C_P is the heat capacity at constant pressure for the core, $q_a = q_d - q_t$ is the power flux available to power the dynamo, q_d is the power flux dissipated at the core mantle boundary, q_t is the threshold power flux for dynamo generation, and c is a constant of proportionality (22, 23). The major difficulty is that the tiny radius of the lunar core, R_c [estimated to be only ~0.2 R (24)], means that core dipole fields are highly attenuated at the planetary surface, given their $(R_c/R)^3$ dependence. Using R_c = 250 to 450 km (see below) and otherwise assuming approximately Earth-like values scaled to lunar core pressures for the remaining parameters, B is predicted to be below 12 μT, and most likely no more than ~3 μT, for essentially all estimated power sources (see table S5). This is well below many of the paleointensity values inferred from the Apollo-era data for the period ~3.6 to 3.9 Ga (Fig. 1).

Furthermore, the data seemed to suggest that lunar paleointensities increased rapidly during the period ~4.0 to 3.9 Ga, such that the field perhaps did not originate until more than ~400 My after lunar formation. It has been argued that the apparent lack of an early dynamo conflicts with the giant impact hypothesis for the origin of the Moon, which would have likely produced a hot, rapidly cooling lunar interior just after formation (3). Moreover, until very recently, essentially all lunar evolution models were unable to explain why the dynamo was apparently able to persist well beyond ~4.1 Ga (25–27).

Another unresolved issue was when the dynamo declined. Fuller (28) argued that the dynamo ceased abruptly sometime between 3.6 and 3.7 Ga, implying that younger paleointensity values are within error of zero field, while Runcorn (3) proposed that the dynamo declined in intensity to ~1 to 20 μT but persisted until at least ~3.2 Ga and possibly much longer. Particularly problematic for the latter viewpoint are the youngest paleointensity values, which are derived from impact melt samples that formed from 1.5 Ga to as little as <200 Ma. If these are accurate, the lack of a field today [<0.2 nT at many surface locations (29)] would seem to suggest that we are observers in a special time window occurring just after the decay of the dynamo.

These issues collectively led to a suspicion that the lunar paleointensity data may not in fact be

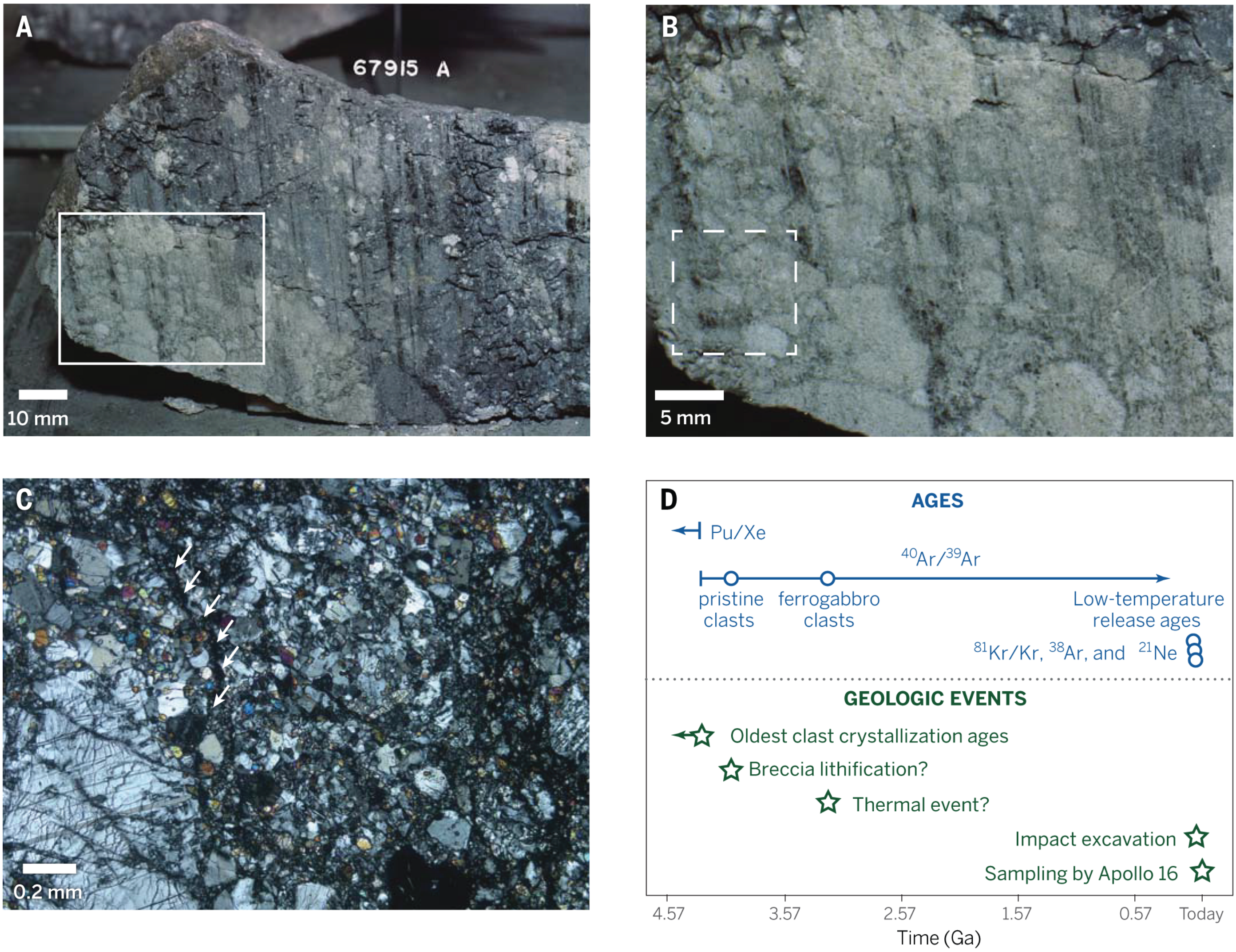

Fig. 2. Petrography and geologic history of Apollo 16 feldspathic polymict breccia 67915. Apollo-era paleomagnetic measurements of this rock, which inferred near-null (<3 μT) paleointensities at ~4.0 Ga, were interpreted as evidence that the initiation of the lunar dynamo was delayed by 500 My after the Moon's formation. However, the magnitude and age of this paleointensity constraint are highly uncertain, given the protracted and complex history experienced by the rock and its constituents. (**A**) Photograph of slab cut at NASA Johnson Space Center in 1972 (NASA photo S72-52255). Numerous diverse clasts are visible, including large (50 by 20 cm) white clast at lower left. Close-up of boxed region is shown in (B). (**B**) Close-up image of large white clast in (A). The large clast is actually itself a polymict breccia composed of numerous rounded white clasts. The thin section shown in (C) was subsampled from below the approximate location shown by the dashed box. (**C**) Photomicrograph in transmitted light with crossed polars of thin section 67915,76 taken from a single rounded white clast at the location shown in (B) (JSC photo 02973-x2). The rounded clast is revealed to be itself a breccia composed of shocked fragments of plagioclase (gray) and pyroxene and olivine (brown, green, and purple grains). The rock is also crosscut by fine-scale (<0.1 mm thick) young, glassy melt veins (dark, sinuous, branching channels) that postdate assembly of the breccia (one example identified with white arrows). (**D**) Summary of radiometric and exposure ages (top) and inferred geologic events (bottom) for 67915. The Pu/Xe and oldest $^{40}Ar/^{39}Ar$ ages indicate that the oldest materials crystallized at ≥4.3 Ga. $^{40}Ar/^{39}Ar$ age spectra from a diversity of clasts—including granulitic breccias, pristine troctolites, and sodic ferrogabbros—exhibit a range of apparent ages from 4.3 to <0.26 Ga, thought to correspond to multiple thermal events recorded by clasts with different Ar diffusivities. $^{81}Kr/Kr$, ^{38}Ar, and ^{21}Ne cosmic ray exposure ages are thought to date the impact that formed Cone Crater and excavated 67915. Ages and paleomagnetic data are presented in table S7.

a record of an ancient dynamo. The young paleointensity values, all of which are associated with impactites, motivated an alternative hypothesis that some or perhaps even all lunar paleofields were the products of plasmas transiently generated by meteoroid impacts on the lunar surface (*30*, *31*). Theoretical models have suggested that such fields might reach tens or even hundreds of μT but would be short-lived, lasting up to a day for the largest, basin-forming impacts and less than seconds for most smaller impacts. Although impact fields would only be capable of imparting a full TRM on samples that cool from the Curie temperature to surface temperatures more quickly than these time scales (*32*), an SRM could be readily acquired by virtually any rock shocked in the presence of such fields, given the near instantaneous pressure changes associated with shock waves (*33*). If correct, the impact plasma fields hypothesis would have profound implications for all of extraterrestrial paleomagnetism, because it implies that NRM may not be a record of internal geophysical processes on planetary bodies.

The impact plasma fields proposal appeared to be critically bolstered by Apollo subsatellite magnetometry, which mapped the lunar field between latitudes of 35°N and 35°S and identified numerous crustal magnetic anomalies (*34*). These anomalies can be interpreted as either localized crustal regions with anomalously high χ_{TRM} (resulting, for example, from high metal abundances) and/or regions exposed to locally intense magnetizing fields. The origin of the anomalies was enigmatic because they did not appear to have clear associations with geological features. A major exception was the most spatially extensive and intense anomalies, which are located on the southern lunar farside in a region that is approximately antipodal to the four youngest large (>600 km in diameter) lunar basins: Imbrium, Serenitatis, Crisium, and Orientale (Fig. 3). This association led some investigators to interpret this association as the result of magnetic field amplification (i.e., locally high B_{paleo}) by impact-generated plasmas in these regions (*34*). Regardless of the formation mechanism, the southern farside and other strong anomalies indicate that large volumes of the crust must have extremely high mean NRM: Assuming a 1-km-thick source layer, the NRM is ~1 to 10 A m^{-1}, which is ~1 to 3 orders of magnitude larger than the NRM of most Apollo mare basalts and breccias and 3 to 5 orders of magnitude larger than the NRM of pristine feldspathic rocks, which are thought to make up the majority of the lunar crust (*35*).

By the end of the Apollo era, it was certain that there were magnetic fields on the ancient Moon, but the intensity, timing, and, most important, physical origin of these fields were unclear. Whether the magnetization in lunar materials was the product of a core dynamo

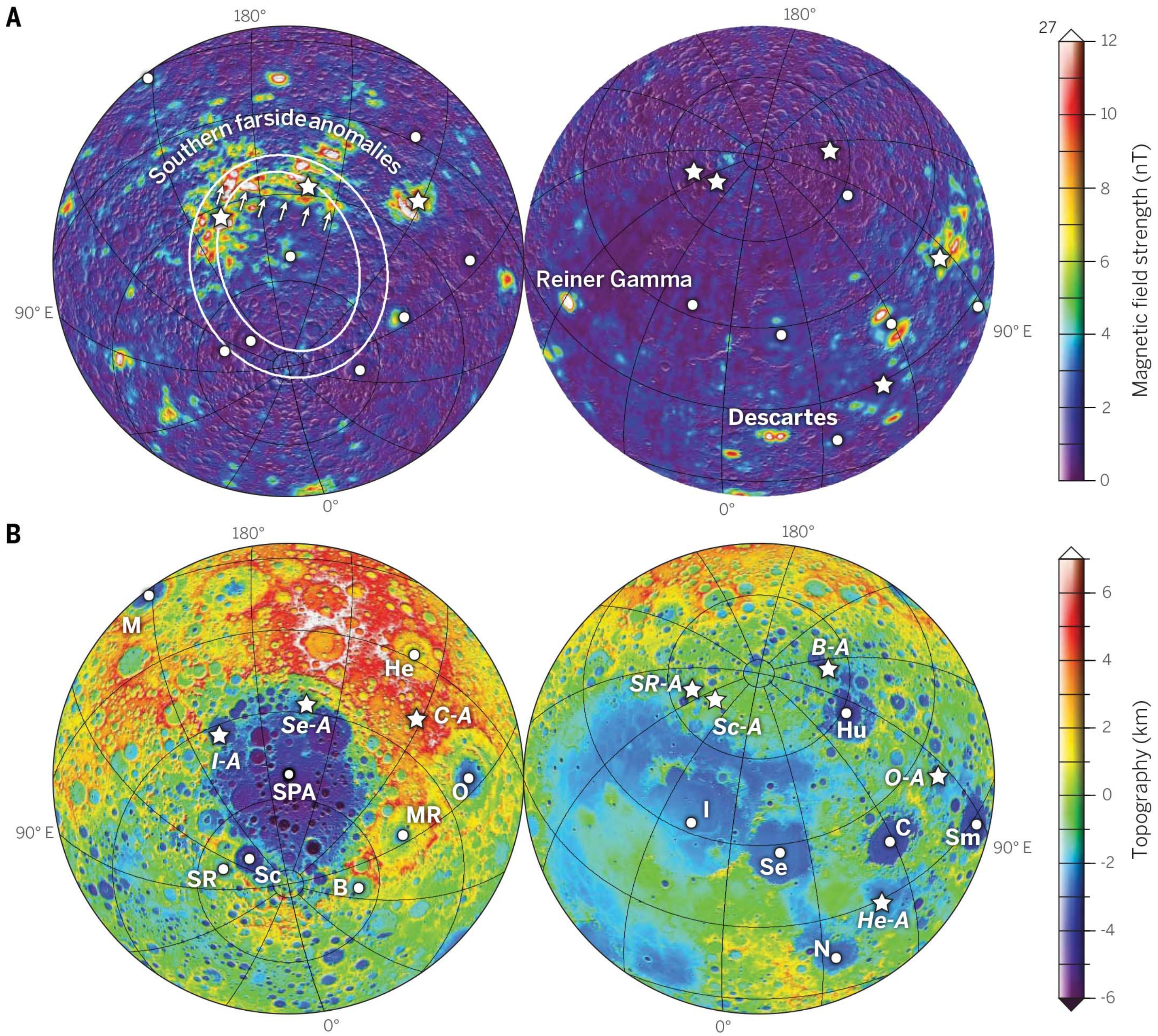

Fig. 3. Lunar crustal magnetic anomalies and their geologic relationships. (**A**) Lunar Prospector total magnetic field map at 30-km altitude from the sequential model of (*8*). Prominent magnetic anomalies (southern farside, Reiner Gamma, and Descartes) are identified. Nested ellipses denote the inner basin floor and outer structural rim of the SPA basin (*82*). Arrows denote a possible linear dike, as identified by (*71*). Circles and stars represent impact basin centers and impact basin antipodes identified in (B). (**B**) Topography from the Lunar Orbiter Laser Altimeter (*83*), showing centers of selected major impact basins (circles) and the antipodes of the eight youngest impact basins (stars). Basin centers are identified by the following: B, Bailly; C, Crisium; M, Moscoviense; MR, Mendel-Rydberg; He, Hertzsprung; Hu, Humboldtianum; I, Imbrium; O, Orientale; N, Nectaris; Sc, Schrödinger; Se, Serenitatis; Sm, Smythii; SPA, South Pole-Aitken; and SR, Sikorsky-Rittenhouse. Basin antipodes are identified with the previous abbreviations appended by "–A". [Modified from (*72*)]

or impact-generated plasmas remained unresolved. A key difficulty was that criteria for distinguishing between these two field sources from measurements of NRM had not been firmly established.

Modern developments

Since the end of the Apollo era, there have been major improvements in analytical techniques in geochronology, petrology, and paleomagnetism and a deepening understanding of rock magnetism and global lunar geophysics. Beginning about 6 years ago, a new generation of paleomagnetic analyses, thermal evolution models, and dynamo simulations have applied these advances toward resolving the origin of lunar magnetism. The broad goal of these efforts has been to answer the following questions: Did the Moon form a dynamo? If so, how strong was the dynamo? When did it form? When did it decline? And what were the physical field-generating mechanisms and power sources? We begin by discussing recent theoretical advances in our understanding of the lunar dynamo and then compare their predictions to new magnetic measurements of Apollo samples and the lunar crust.

Lunar dynamo mechanisms

A dynamo requires organized advection of conducting fluid, which generates magnetic fields by the process of electromagnetic induction (*36*). For the terrestrial planets, this is thought to occur in their liquid metallic cores. Therefore, if the Moon once had an internally generated, global magnetic field, this implies the existence of an ancient advecting lunar metallic core. The case for the existence of such a core has been greatly strengthened over the past two decades. Lunar Prospector estimates of the lunar moment of inertia (*37*) and induced dipole moment (*38*, *39*) are consistent with a core of radius ~220 to 450 km. Lunar rotational dissipation inferred from four decades of laser ranging further suggests that the core is at least partially in a fluid state (*40*). Recent reprocessing of Apollo seismic data suggests the presence of a ~250- to 420-km radius liquid Fe outer core possibly surrounding a ~200- to 240-km radius solid inner core (*24*, *41*).

Fluid motions in all known dynamos in the solar system today are generally thought to be powered by convection. The buoyancy flux in Earth's core is provided by secular heat flow, as well as by latent heat release and compositional differences arising from crystallization of the core (*42*). The field's intensity (*22*) and longevity (*36*) are thought to depend on the heat flux at the core-mantle boundary (Fig. 4). The heat flux out of the core is in turn strongly controlled by the thermal evolution of the overlying mantle. Purely thermal convection in the core (i.e., occurring in the absence of core crystallization) requires that the core-mantle boundary heat flux exceed that which can be conducted along a core adiabat. Estimates for the lunar core adiabatic heat flux range widely (between ~2 and 10 mW m^{-2}) (*43*).

For a dry lunar interior, thermal evolution models assuming a monotonically cooling core-mantle boundary typically find that a pure thermal convection dynamo can persist at most until sometime between ~4.3 and 4.1 Ga (*25*, *26*, *44*). Enrichment of the lower lunar mantle by 40 parts per million (ppm) H_2O [below that recently inferred from mare basalts (*45*)] may lower the mantle viscosity by ~2 orders of magnitude, thereby enabling a thermal convection dynamo to persist until perhaps ~3.4 Ga (*46*). Additionally, it is possible that at the end of magma ocean crystallization, the core was surrounded by a radiogenic element-rich cumulate layer. Thermal models (*26*) have estimated that radioactive heating from such a layer could delay initiation of a thermal convection dynamo (*47*) until sometime between 4.2 and 4.0 Ga and postpone the dynamo cessation until perhaps ~3.5 Ga. The generally short lifetime predicted for a purely thermal convection dynamo has motivated alternative hypotheses that the lunar dynamo was powered either by thermochemical convection from core crystallization or mechanical stirring by differential rotation between the mantle and core.

Mechanical dynamos have yet to be definitively identified to operate in any planetary body today. At least two forms of these exotic mechanisms have been described for the Moon. The first proposes that large meteoroid impacts, by driving transient large-amplitude mantle librations or even temporarily unlocking the mantle from synchronous rotation, powered dynamos lasting for up to several thousand years after each impact event (*48*, *49*). However, because this mechanism requires highly energetic impact

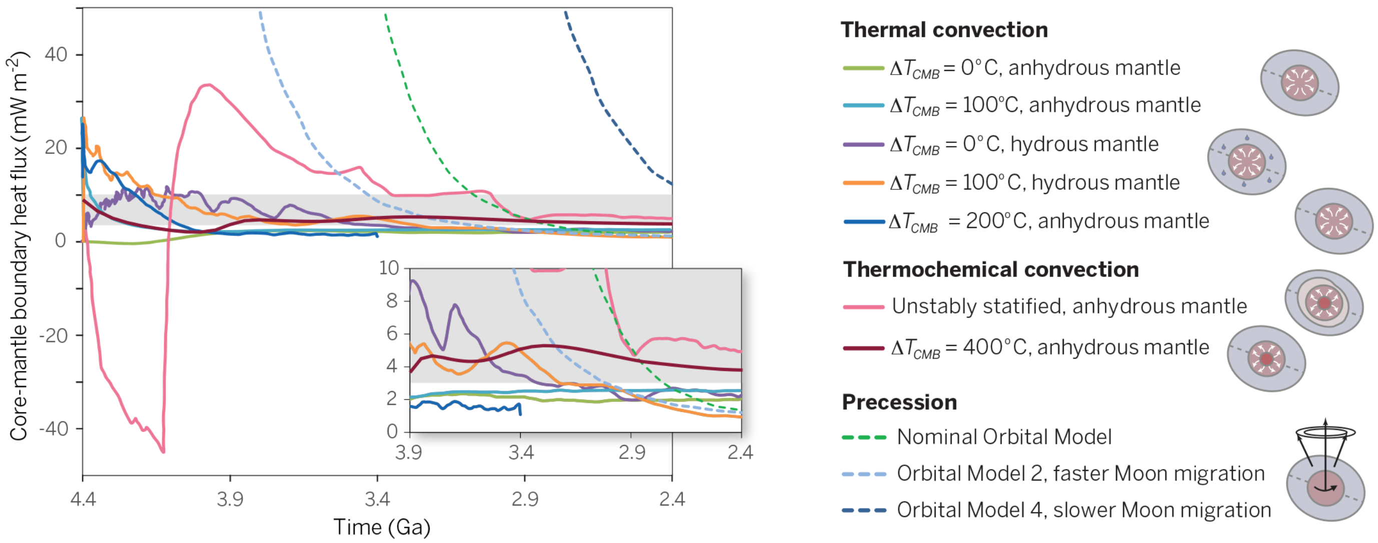

Fig. 4. Estimated longevity of various lunar dynamo mechanisms versus time. Shown is the estimated core-mantle boundary energy flux supplied by purely thermal core convection, thermochemical convection driven by core crystallization, and precession. Solid curves depict a variety of end-member convective dynamo power models demonstrating the effects of: (i) an initially stably stratified (all curves except light red curve) or unstably stratified (light red curve) mantle; (ii) a completely anhydrous (all curves except purple and orange) or hydrous (40 ppm H_2O) (purple and orange curves) mantle; (iii) an initial temperature difference between the core and mantle (ΔT_{CMB}) of either 0°C (green and purple curves), 100°C (light blue and orange curves) [models D07, D09, D37, and D39, respectively, of (*46*)], 200°C (dark blue curves) [model TB-0 of (*26*)], or 400°C (dark red curve); and (iv) inner core crystallization (light red and dark red curves) [model H50E100V5e19 of (*54*) and ΔT_{CMB} = 400 K, 4 wt % S model of (*56*), respectively]. Dashed curves depict estimated power from mantle precession for the three orbital models considered by (*23*): Nominal (green), Model 2 (light blue), and Model 4 (dark blue). Gray boxes indicate the estimated range of critical core heat flux values required for field generation in thermal convection dynamos (3 to 10 mW m^{-2}). Dynamos driven by precession and compositional convection may operate subadiabatically. (Inset) Magnified model results for the time spanning 3.9 to 2.4 Ga. Because the models shown here assumed modestly differing properties of the lunar interior (e.g., core radius and sulfur abundance), their results should only be compared qualitatively.

events, it seems only likely to have occurred before the final basin-forming impact that formed the crater Orientale at 3.72 Ga (50).

A mechanical dynamo could also be powered by mantle precession (40, 51). Although initially locked, the lunar core and mantle began to precess around different axes when the Moon's orbit migrated to beyond >26 to 29 Earth radii (52). The power associated with this differential motion should have been extremely large compared to that driving thermal convection (Fig. 4) but declined over time as the angle between the core and mantle spin axes decreased. The lifetime of the precession dynamo is highly uncertain because it could operate subadiabatically and because the lunar orbital evolution is poorly constrained. Nevertheless, for a typical choice of lunar physical parameters (23), it is estimated that such a dynamo could be sustained until the Earth-Moon distance reached ~43 to 48 Earth radii (i.e., until ~3.4 to 2.0 Ga, depending on which lunar orbital evolution model is assumed). Considering a broader range of acceptable lunar parameters, it is conceivable that a precession dynamo could even last until as late as ~0.6 Ga (14). Not only is a precession dynamo potentially very long-lived, but very recent magnetohydrodynamic (MHD) simulations (53) also have found that it might have produced surface fields up to the ~100 μT intensities suggested by the Apollo-era paleointensity record for the period before ~3.6 Ga (Fig. 1). Interestingly, this MHD model also finds that the dynamo would terminate at about the same time that the Apollo-era paleomagnetic data suggest a decline in paleointensities (i.e., after ~3.6 Ga).

None of the above mechanisms currently appear capable of powering a dynamo beyond ~0.6 to 2 Ga. However, they do not take into account the main power source for Earth's dynamo, core crystallization. Thermal evolution models predict that a solid core should have formed within the Moon for core sulfur contents of less than ~12 weight percent (wt %) (44, 54). Given possible seismic evidence for a solid inner core (24) and recent modeling of lunar mantle siderophile element depletions suggesting a bulk core sulfur abundance of ~6 wt % (55), core crystallization could have played an important role in sustaining a late lunar dynamo.

Two lunar thermal evolution models (54, 56) have recently emerged that consider the effects of core crystallization on the dynamo (Fig. 4). However, only one of these models (56) considered the central advantages of core crystallization for prolonging the dynamo: the ability to drive convection even for subadiabatic heat core–mantle boundary heat fluxes and the high efficiency by which these heat fluxes are transmitted into buoyancy fluxes by compositional convection. Given the uncertainties on the core sulfur content, it was found that a dynamo could be powered for billions of years and perhaps even until the present day.

However, core crystallization again appears only able to generate surface fields of just a few μT or less and is therefore apparently unable to explain the high paleointensities from the period before 3.56 Ga. Furthermore, the longevity

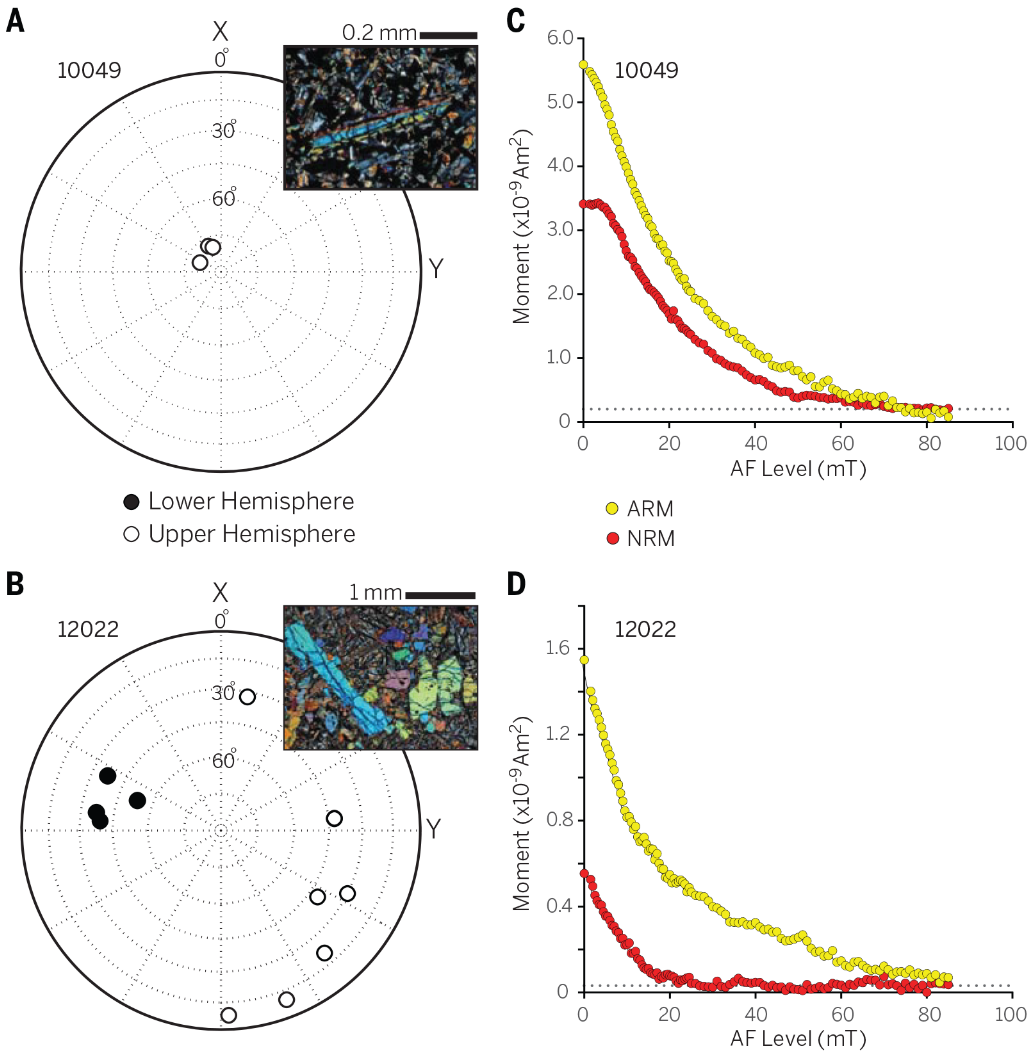

Fig. 5. NRM of two mare basalts that formed during and after the high-field epoch. (**A** and **B**) Equal-area stereographic projection showing the most stable magnetization component directions observed for mutually oriented subsamples of 10049 (A) and 12022 (B). Insets contain transmitted light, crossed polars photomicrographs of thin sections of each rock (10049,40 and 12022,110), showing their unshocked and unbrecciated textures (JSC photos 04194-x3 and 04491-x2, respectively). Visible are plagioclase (gray), pyroxene and olivine (brown, green, and purple grains), and oxides (opaque grains). The sizes of these plagioclase grains indicate that these rocks cooled slowly relative to putative impact-generated magnetic fields (**C** and **D**) AF demagnetization of 10049 subsample 102-1 (C) and 12022 subsample 310a (D). Shown is the NRM intensity during AF demagnetization compared with that of a laboratory ARM, a TRM proxy. Horizontal dashed lines indicate the magnetic moment at which the sample was effectively completely demagnetized (as indicated by instability of the magnetization direction). For 10049, the unidirectional NRM (A) and its persistence to similar AF levels as reached by the ARM (C) are consistent with a primary TRM origin (Box 1). For 12022, the nonunidirectional orientation of the NRM (B) and its complete removal at AF levels lower than that for ARM (D) are inconsistent with a primary TRM origin. Rather, the stable magnetization in 12022 is interpreted to be an overprint from saw-cutting and long-term exposure to Earth's field. [Data from (14, 50)]

of power from core crystallization may open a new problem of explaining why there is apparently no lunar dynamo today. A possible solution is that the core crystallization models discussed here assumed that the lunar core began by crystallizing from the bottom up like the Earth's core. In fact, early lunar core conditions were likely close to the threshold that could permit bottom-up or top-down (Fe snow) crystallization (which depends critically on the poorly known sulfur content of the liquid core and thermal expansion of iron-sulfur liquids) (57, 58). Given that the Moon presently lacks a detectable dynamo field, it was proposed that as the liquid core became progressively enriched in sulfur, a shift from bottom-up to top-down crystallization occurred that induced dynamo shutoff. However, the expected timing for such a transition remains unknown (56).

In summary, it is possible that a lunar dynamo powered by core crystallization could have persisted for far longer than permitted by any other mechanism and for as long as the Apollo-era data seem to require. However, only precession dynamos currently appear capable of accounting for the high apparent paleointensities measured during the Apollo era for the period before ~3.6 Ga. We now reassess the fidelity of this paleointensity record in the light of modern laboratory and spacecraft magnetometry measurements.

Box 1. Sample criteria for lunar dynamo studies.

Lunar dynamo paleointensity studies should target samples containing a high-fidelity TRM acquired in a temporally stable field at a well-constrained time in lunar history. This motivates the following selection criteria:

1. TRM origin. Current experimental procedures are most accurate for inferring paleointensity estimates from samples containing TRM. Samples containing total TRM (i) are igneous in origin and/or were at least heated above the Curie temperature during the last remagnetization event and (ii) show no petrographic evidence for shock (pressures <5 GPa) or brecciation after the last major remagnetization event. To help ensure that their NRM has high stability and can be cleaned of overprinting magnetization, samples should also (iii) contain the finest ferromagnetic grains (ideally pseudosingle domain or smaller). Other key characteristics of total TRM are (iv) unidirectionality within the samples, (v) high stability against laboratory demagnetization, and (vi) demagnetization in the laboratory like a laboratory-produced TRM.

2. Core dynamo field. To ensure that any observed TRM was acquired from a long-lived field like a dynamo, samples should have cooled from 780° to ~0°C more slowly than the estimated lifetimes of impact-generated fields [~1 day for the largest impacts (30) and at most a few seconds for the smaller impacts after 3.7 Ga (14)]. A lack of major shock effects (criterion ii in 1) will also help to exclude an impact-generated field source.

3. Known age. To enable the assignment of accurate ages to paleointensity estimates, samples should have geochronologically constrained thermal histories. Identification of a TRM origin (criteria i to vi in 1) is important for assigning radiometric ages to the NRM.

New paleomagnetic measurements

Modern paleomagnetic studies have taken advantage of new, highly sensitive and automated magnetometry technology and four decades of continued geochemical and petrologic studies of the Apollo samples. A particularly important contribution has been the development of several criteria for identifying high-fidelity dynamo paleointensity records (Box 1). Critically, rocks have been studied that are unbrecciated and unshocked and with cooling time scales from the kamacite Curie point to ambient surface temperatures far longer than any putative impact plasma-generated fields, requiring that any primary TRM be the product of a long-lived field like a core dynamo (Fig. 5).

Such studies have now provided strong evidence for the existence of a lunar core dynamo. Analyses have found that the lunar troctolite 76535 recorded a dynamo field with intensity a few tens of μT at 4.25 Ga (32, 59). This is the earliest record of the lunar magnetic field and demonstrates that a dynamo existed as far back in history as we have records (Fig. 6). This early period is exactly the time when thermal evolution models predict that a purely thermal convection dynamo could have been active. Furthermore, this result marginally excludes the thermal blanket convection dynamo models (26, 54), which predict that dynamo initiation would be delayed until ~4.2 to 3.9 Ga and would perhaps coincide with the onset of major mare volcanism (~4.0 Ga).

Subsequent studies of mare basalt 10020 identified evidence for a core dynamo field with an intensity of ~66 μT at 3.72 Ga (60). Additional evidence for a similarly substantial field at 3.7 to 3.8 Ga was also provided by a study of three Apollo 17 mare basalts (61). The existence of a dynamo at 3.8 to 3.7 Ga likely excludes a purely thermal convection dynamo (25), unless the lunar mantle is H_2O-rich (46) (Fig. 6).

The lifetime of the dynamo was then found to persist until at least 3.56 Ga, with similarly strong intensities (~71 to ~77 μT) from analyses of mare basalts 10017 and 10049 (Fig. 5, A and C) (50). Given that these samples are 160 My younger than the last basin-forming impact, they likely exclude an impact-driven mechanical dynamo at this time in lunar history in favor of a precession, core crystallization, and/or thermal convection dynamo in an H_2O-rich lunar interior (Fig. 6).

After these studies, the question remained as to when the dynamo declined and ultimately ceased. A key recent step toward addressing this has been the realization that the poor magnetic recording properties of most mare basalts preclude retrieval of TRM acquired in fields less than ~20 μT using typical AF demagnetization methods, which introduce spurious magnetization that can obscure the underlying NRM (62). This limit is barely within the range of the Apollo-era paleointensity values after 3.5 Ga, strongly suggesting that most of those values are simply upper limits and are consistent with a weak to null field. Therefore, establishing the history of the late lunar dynamo requires measurements of samples with unusually high magnetic recording fidelity. Recent analyses of two such samples, mare basalts 15597 and 12022, demonstrated that the dynamo field had declined to below ~7 μT by ~3.3 Ga and below ~4 μT, and possibly even vanished altogether, by 3.19 Ga (Fig. 5, B and D, and Fig. 6) (14). These values are broadly consistent with a possible 5-μT upper limit obtained for Thellier-Thellier measurements of <3.46 Ga anorthositic breccia 60015 (14, 18) (Fig. 6). Finally, a recent paleomagnetic study of a <7-Ma impact melt glass demonstrated that lunar surface fields likely remained below ~7 μT into the recent past (62, 63) (Fig. 6).

The ~4-μT upper limit inferred for the Moon at 3.19 Ga is nevertheless at the upper end of the range predicted by the energy flux scaling (Eq. 2). In particular, for typically assumed lunar parameters, essentially all published thermal and evolution models of the lunar dynamo (23, 26, 44, 47, 48, 56) predict surface fields >0.1 μT for more than ~90% of the time period in which the dynamo is active in the models. Such a minimum field is comparable to estimates of the strongest lunar crustal surface fields (64) and below even the weakest known dynamo surface field in the solar system today (65). Therefore, the current paleointensity constraints do not at all exclude a lunar dynamo after 3.56 Ga. Determining when the dynamo truly ceased requires a sample with extraordinary magnetic recording capabilities. Tikoo *et al.* (66) very recently suggested that the melt-glass matrix of regolith breccia 15498 (which contains a population of high-fidelity single-domain ferromagnetic grains) was magnetized in a ~2-μT core dynamo field. The age of the TRM is uncertain but must be less than the 3.3 Ga age of basalt clasts in the breccia, whereas trapped Ar data suggest a lithification age of perhaps only ~1.3 Ga (67). If confirmed, this result would extend the lifetime of the lunar dynamo by at least 0.2 to 2 billion years, suggesting that the dynamo operated in two different intensity regimes: a pre-3.3-Ga high-field epoch and a post-3.3-Ga weak-field epoch. Depending on the age of the NRM in 15498, precession and core crystallization are the only mechanisms currently thought

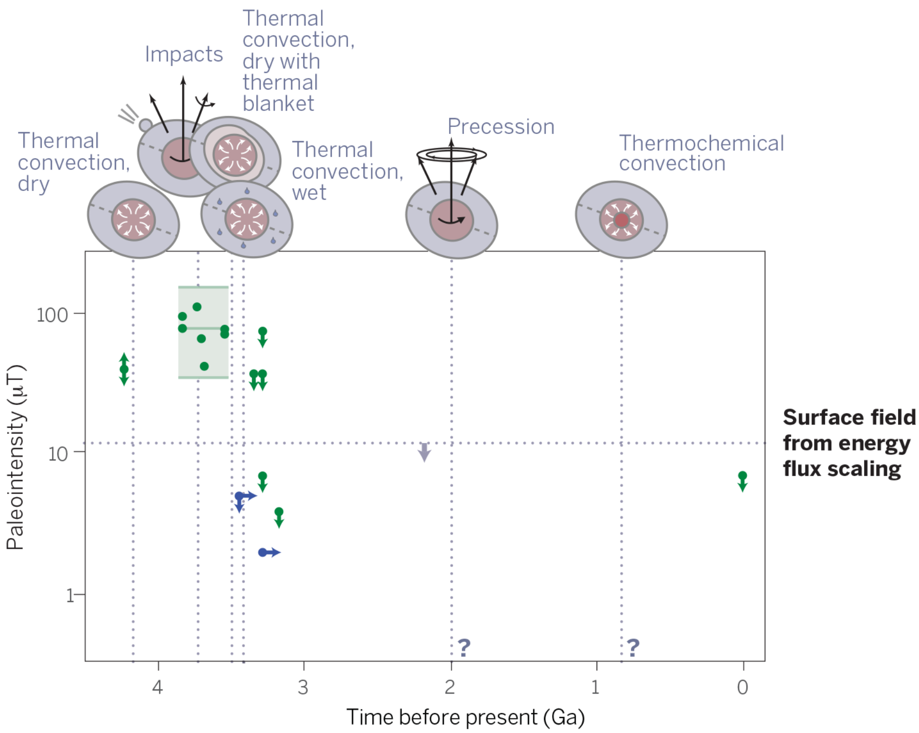

Fig. 6. Paleointensity measurements of the lunar magnetic field using modern methods and estimated lifetimes of various lunar dynamos. Each point represents measurements of a single Apollo sample. Circles represent actual paleointensities, downward arrows represent upper limits on paleointensity, and right arrows represent upper limits on age. Green and blue points were measured using the IRM and Thellier-Thellier methods, respectively, and the upper limits were derived from the ARM method. Note the datum at <7 Ma at the extreme right of the figure. The shaded green box encompasses the mean paleointensity value for the period 3.56 to 4.25 Ga (central green line) and its estimated 2 SD uncertainty (upper and lower lines) (see table S6). The paleointensity value for 76535 (leftmost green point) is currently not well constrained due to spurious demagnetization effects. Vertical dashed lines show estimated maximum lifetimes of various proposed lunar dynamo mechanisms: purely thermal convection in a dry mantle (with and without an early thermal blanket surrounding the core); impact-driven mantle rotational changes, purely thermal convection in a wet mantle; precession; and thermochemical core convection driven by core crystallization. The horizontal line shows the maximum lunar surface field as estimated from Eq. 2. The lifetimes of the precession and thermochemical dynamos are highly uncertain. Paleointensity data set is from (*14*, *18*, *50*, *59*–*61*, *63*, *66*) and is also listed in table S6.

to be able to power a dynamo this late in history (Figs. 4 and 6).

These results confirm some Apollo-era conclusions and refute others. Modern paleomagnetic measurements have now demonstrated that a core dynamo likely existed on the Moon at least as far back in time as 4.25 Ga, in contradiction to the hypothesis of a late origin (i.e., after 4.0 Ga). On the other hand, modern measurements have confirmed the existence of a high-field (mean of ~77 μT from six samples) epoch lasting from at least 3.85 to 3.56 Ga, as originally recognized during the Apollo era. However, we now see that the dynamo subsequently declined precipitously to <~4 μT by 3.2 Ga. In particular, modern analyses of samples 3.3 Ga or younger suggest that most or all of the Apollo-era paleointensity analyses from this late epoch, which ranged up to apparent values of 20 μT, are only upper limits on the lunar field and therefore do not require a dynamo at this time. Finally, a recent analysis of a potentially very young lunar sample (*66*) hints that the dynamo may have nevertheless persisted in a weakened state (surface field of ~2 μT) until sometime after 3.3 Ga.

The global context of crustal magnetism

Coincident with these new paleointensity experiments, there have also been important recent investigations of the lunar crustal magnetic anomalies. As discussed above, a key finding from the Apollo-era measurements was the discovery of several intense magnetic anomalies, including those on the southern farside, that apparently require crustal materials with NRM larger than that of nearly all known Apollo samples (*35*). Global magnetic field maps from the Lunar Prospector spacecraft, which only became available 6 years ago (*8*, *29*, *68*), have now confirmed the unique size and intensity of the southern farside anomalies and their spatial correlation with antipodes of four, or perhaps even five (*69*), of the eight youngest basins. However, the acquisition of the first high-resolution global topography data in 1993 by the Clementine laser altimeter (*70*) led to the identification of another geologic association with the southern farside anomalies: They lie on the northern edge of the vast South Pole-Aitken (SPA) basin, the largest known impact crater on the Moon.

This discovery has led to two explanations for these anomalies as alternatives to the impact-plasma fields hypothesis. Both propose that rather than reflecting locally high B_{paleo}, they are a manifestation of materials with locally high χ_{TRM} that were magnetized in a dynamo field. The first of these proposals (*71*) identifies the anomalies with the inferred locations of <4.0 Ga basaltic dikes located within the SPA (Fig. 3). The assumed great vertical thickness (30 km) of these dikes enables them to produce the strong SPA anomalies despite the weak NRM of typical mare basalts.

The second proposal (*72*) emphasizes that the anomalies are in the expected locations of ejecta and impact melt derived from the SPA impactor (Fig. 3). The key advantage of this proposal is that typical chondritic impactors, which are far more Fe-rich, have χ_{TRM} ~100 and 10^4 times that of mare basalts and pristine feldspathic rocks, respectively. Therefore, deposition of relatively thin layers (perhaps just tens to thousands of meters thick) of such material should have fundamentally enhanced the magnetic properties on the surface. SPA ejecta, as well as that from more recent impacts, could also account for most of the other isolated crustal anomalies. However, thus far no evidence has yet been identified indicating that at least the surface layer (top ~30 cm) of the southern farside anomaly region is anomalously iron metal-rich (*73*).

Both of these new proposals require only dynamo fields rather than plasma fields to explain the magnetic anomalies. Yet more evidence for a dynamo comes from the recent identification of many anomalies within the interiors of most Nectarian (i.e., ~3.85 to 3.92 Ga) basins and some pre-Nectarian (i.e., more than ~3.92 Ga) basins (Fig. 3) (*74*–*76*). These impacts likely demagnetized any existing NRM, producing impact melt sheets that should have cooled from the kamacite Curie point to ambient lunar surface temperatures far more slowly (given their >1 km thicknesses, over a period of >10^4 years) than any impact-generated field. Therefore, these central basin anomalies likely require a steady field like that expected from a core dynamo during the Nectarian and pre-Nectarian periods. However, once again, the inferred TRM intensities (~1 A m^{-1}) exceed what is achievable with known endogenous lunar materials and may reflect the addition of Fe-rich impactor materials.

It has been suggested (*29*) that the apparent lack of anomalies in the centers of Imbrian basins (<1 nT at the surface, suggesting that the central impact melt layer has an NRM less than ~10^{-3} A m^{-1}) may indicate that the dynamo was no longer active by the time of their formation (i.e., after ~3.85 Ga). However, this proposal is not consistent with the strong evidence for a core dynamo from paleomagnetic analyses of 3.56 to 3.85 Ga (Imbrian-aged) Apollo 11 and 17 basalts (see above). In fact, the sample and orbital data sets are consistent if the composition of the impact melt in these locations is similar to that of pristine feldspathic rocks or if the dynamo reversed during the cooling time of the impact melt (perhaps several My, assuming a 10-km thickness). Therefore, the lack of central basin anomalies does not exclude the existence of a dynamo during the Imbrian epoch. The strong anomalies observed in the older Nectarian basins (see above) could still be explained by impact melts with higher χ_{TRM} (although it is unclear why these would only occur in the older basins) or perhaps a longer reversal period during this earlier time.

Key unknowns about the lunar field

The new data make it clear that a dynamo once existed on the Moon. A central remaining mystery is now the nature of the physical mechanism(s) that powered it. The challenge is to find a dynamo-generation mechanism that can account for several outstanding features of the lunar paleointensity record. Foremost is the extremely high paleointensities (average of 77 μT) inferred for the period 3.56 to 3.85 Ga (table S6). Even though we estimate that these individual paleointensities each likely has a 2-SD uncertainty of a factor ~5 (tables S2 to S4), the mean value should only have a 2-SD uncertainty of factor ~2.2 if the samples formed in a field of constant intensity (table S6). Therefore, we consider that any successful dynamo model must predict a minimum paleointensity of ~35 μT for the high-field epoch, at least three times the maximum value estimated by the energy flux scaling (Fig. 6).

One possibility is that the lunar dynamo field strengths are underestimated by Eq. 2. This is conceivable given that this relationship was inferred from convective dynamo models only and, moreover, that these models operate in a highly viscous regime not found in planetary cores (*77*, *78*). Given the possible recent prediction by a precession MHD model of paleointensities up to 100 μT (*53*), this seems to be likely for at least some mechanical dynamos. Another less likely possibility is that Eq. 2 is applicable to the lunar dynamo, but one or more of the assumed parameter values are inaccurate. Although the values of many of these parameters are indeed uncertain, the form of Eq. 2 shows that B is only strongly dependent on R_c/R (and to a lesser extent, on d). For example, even simultaneously lowering the adiabatic power threshold to its minimal value of 0 W m^{-2} while increasing the dipolarity d to its maximal value of 1 would be insufficient to explain the paleointensities during the high-field epoch. Increasing the core size to an Earth-like value of about half the lunar radius could account for the high paleointensities, but this appears to be incompatible with the moment of inertia and bulk density data (*12*, *13*).

A related problem is presented by the order of magnitude decline in paleointensities during the period 3.56 to ~3.3 Ga. This is because no single lunar dynamo model has yet been able to account for such a rapid decline while simultaneously enabling the dynamo to persist in a weakened state (if we accept the evidence for a ~2-μT dynamo persisting until sometime between 3.3 and 1.3 Ga from sample 15498). If we assume that this rapid paleointensity decrease was a manifestation of a declining power source, then the energy flux scaling law (Eq. 2) predicts a rapid relative drop in paleointensities only when q_a approaches 0 (i.e., when q_d approaches q_t). Thus, unless the absolute magnitude of the time derivative of q_d rapidly dropped after 3.56 Ga, a given power source is likely to become insufficient to power dynamo action (i.e., to maintain $q_d > q_t$) for another 0.2 to 2 billion years after the paleointensities declined. A possible solution is to postulate at least two distinct dynamo mechanisms, one during the high-field epoch [perhaps with field intensities described by a relationship distinct from Eq. 2, as may be the case for the precession dynamo (*53*)], and a different dynamo mechanism afterwards. This is appealing because the apparent ~2-μT paleointensities during the later history of the dynamo are well predicted by scaling relationships for the core crystallization dynamo scenarios in late lunar history (*56*) (Fig. 6). A second speculative explanation is that only a single dynamo mechanism acted over the entire period from 4.2 to 1.3 Ga but that this dynamo was bistable, transitioning from a strong, dipole-dominated geometry to a weaker, multipolar state between 3.56 and 3.3 Ga. Bistable dynamos have been observed in dynamo models when the local Rossby number, which quantifies the ratio of inertial to Coriolis forces, is less than ~0.1 (*79*, *80*). Although the nature of the physical events that would have caused the Moon to transition from the strong to weak branch is uncertain, such transitions have been observed to occur spontaneously after reversals (*81*). Paleomagnetic records constraining the geometry of the lunar paleofield and its evolution in time would be an important test of this mechanism.

A final problem is the persistence of the lunar dynamo. The new evidence for a dynamo lasting until at least 3.3 Ga, and perhaps as late as 1.3 Ga, makes extreme demands on the longevity of the power source. At present, only thermochemical convection driven by core crystallization can supply such long-lived power, but that may open a new problem of explaining how the dynamo subsequently turned off.

Outlook

A diversity of geophysical and geochemical evidence now indicates that the Moon formed a ~1 to 4 wt % metallic core. This core was once advecting and generated a dynamo magnetic field. As far back in time as we have paleomagnetic records (4.25 Ga), it appears that the Moon had a dynamo, consistent with an early, partially molten body like that expected after a giant impact origin. The Moon is therefore a highly differentiated object like the terrestrial planets.

The surface field intensity was in the range of several tens of μT from at least 4.25 to 3.56 Ga but then declined precipitously (by at least an order of magnitude) between 3.56 and 3.19 Ga. The only mechanism that has yet been proposed to be capable of generating such high intensities in the early period is precession. The next major steps in the magnetic exploration of the Moon should be to obtain better constraints on global geophysical parameters (particularly the adiabatic heat flux and sulfur content of the core and a better understanding of the thermal evolution of the mantle) and accurate measurements of the field intensity in time. Furthermore, a determination of the time when the dynamo finally ceased would have major implications for constraining the power mechanism for the late dynamo. Eventual measurements of the absolute orientation of the lunar magnetic field from laboratory measurements of oriented samples and in situ measurements of lunar bedrock should provide transformative constraints on the geometry of the field, its time variability (reversal frequency and secular variation), and local and global-scale tectonics like crustal deformation and true polar wander.

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16. There have been a number of attempts to recover the absolute paleodirections of lunar fields from Apollo samples using analyses of magnetic anisotropy fabrics and from the lunar crust using spacecraft field maps (table S1). However, the paleohorizontal indicators in the Apollo samples are indirect and analysis of the spacecraft data are limited by the fundamental nonuniqueness of inferring magnetization from

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SUPPLEMENTARY MATERIALS

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