which come from beyond the furthest reaches of the solar system, is so low (Tobias Owen, University of Hawaii, and Kevin Zahnle, NASA, Ames Research Center).

A solution may come from the idea (Takuo Okuchi, Nagoya University) that the volatile budget of Earth has been affected by the dissociation of water, the extraction of large amounts of hydrogen into the core, and the oxidation of the iron in Earth's mantle. An excellent new model has been developed (Don Porcelli, California Institute of Technology) for Xe isotope data in terms of late catastrophic volatile loss about 100 million years after the start of the solar system. Did this happen during a Giant Impact–like event? Perhaps there were several such events and the final damage happened after the moon formed. It is unlikely that the moon did not form until 100 million years after the start of the solar system. If Earth had only half formed at this late stage the accretion rates must have been very slow. Furthermore, the W isotope heterogeneity on the moon would have to be an inherited feature because it can only be produced within the first 60 million years of the solar system. Such inherited heterogeneity is hard to reconcile with the energetics of the Giant Impact.

We have recently come a long way in obtaining hard constraints on the origin of Earth and the moon. The issues have changed from discussion of whether or not there was a giant moon-forming impact to debates about the accretion rates of the Earth and the chemical, isotopic, and physical effects of such castastrophic accretionary scenarios.

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PERSPECTIVES: ORIGIN OF EARTH AND MOON

A Couple of Uncertain Age

Frank A. Podosek

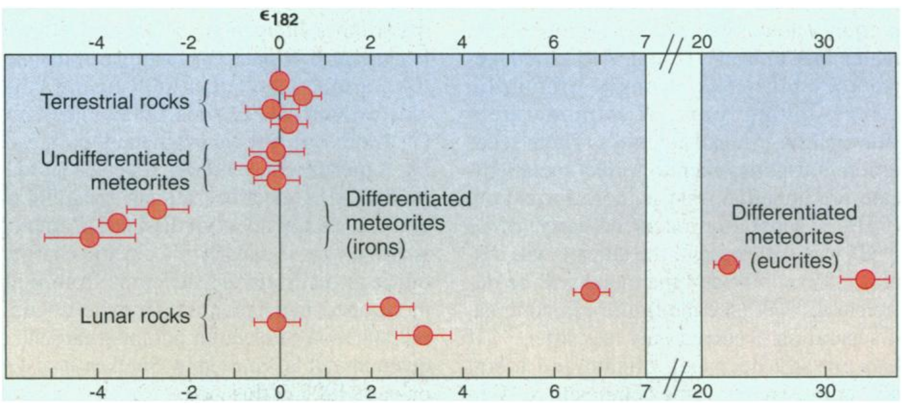
The age of our planet is interesting in its own right and has a bearing on other questions such as early conditions on Earth and the origin of life. In modern practice, geologic ages are determined by measuring the accumulated amount of some daughter isotope that is produced in radioactive decay and relating it to the abundance of the parent radionuclide. In this century, Earth's age has been progressively more constrained as a result of better understanding of natural radionuclide parent-daughter systems and advances in analytical technology. But despite such progress, some thorny issues remain, not only in dating but also in identifying the key processes associated with Earth and moon formation (*1*, page 1861).

The age of the solar system as a whole is easier to determine than the age of Earth. The former is reliably inferred from the age of refractory element–rich inclusions in undifferentiated meteorites to be about 4.57 billion years (*2*), thus providing an upper limit to the age of Earth. These inclusions are the oldest known objects in the solar system, and their content of very short-lived radionuclides such as ^{26}Al (with a half-life of 0.74 million years) indicates that the solar system did not exist for more than about 1 million years before the inclusions formed (*3*).

In contrast to these ancient extraterrestrial objects, there are no known terrestrial rocks or minerals whose formation essentially coincides with formation of Earth, and therefore its age must be inferred indirectly. Several independent approaches indicate that Earth formed about 100 million years later than the solar system as a whole. This is rather a long time compared with theoretical estimates for early solar system evolutionary time scales and for formation of other terrestrial-type planetary bodies, both around 10 million years or less (*4*).

One line of evidence, pursued over several decades, emerges from measurements of the proportions of ^{206}Pb and ^{207}Pb, which are produced at different rates from ^{238}U and ^{235}U, respectively. Models for the evolution of Pb isotopic composition in terrestrial mantle samples almost always give an age of Earth several tens to more than a hundred million years younger than that of the solar system (*5*). Another approach that leads to the same conclusion (*6*) is based on the observation that the amounts of excess ^{129}Xe and ^{136}Xe in the atmosphere are much less than would be expected from early solar system–estimated abundances of their short-lived parent radionuclides ^{129}I and ^{244}Pu, respectively. Similarly, decay of ^{87}Rb for several tens of million years is required to produce Earth's estimated initial ^{87}Sr/^{86}Sr ratio (*7*). The most recent isotopic system applied to the problem is based on the decay of short-lived ^{182}Hf (half-life of 9 million years) to ^{182}W (see the figure); it too suggests formation after several tens of million years (*8*).

The reader will note my frequent use of qualifiers. All of the above-mentioned isotopic chronometers are intrinsically capable of considerably higher precision, but this precision cannot yet be realized. It is not even clear whether the chronometers are consistent or in conflict with each other. One reason is that the issue is not so simple



Tungsten isotope survival. Relative abundance of ^{182}W, the daughter of short-lived ^{182}Hf (half life of 9 million years), in various solar system materials [data from (*8*)]; the abscissa scale is in units of 0.01%. If a planetary body experiences metal-silicate fractionation while ^{182}Hf is still present in substantial amounts, W in the metal should be depleted, and W in the silicate enriched, in ^{182}W. This appears to be the case for meteoritic metal and for lunar and meteoritic silicate but not for Earth's mantle and crust.

The author is in the Department of Earth and Planetary Sciences, Washington University, St. Louis, MO 63130, USA. E-mail: fap@levee.wustl.edu

as just dating a rock. Instead, all methods rely on models of varying complexity involving assumptions difficult to verify and parameters difficult to measure. In addition, radiometric dating is generally formulated in terms of chemical events such as fractionation between parent and daughter elements. In the real history of Earth, the relevant chemical changes likely must be treated as protracted processes rather than instantaneous events. Perhaps most fundamentally, different isotopic systems record different chemical changes, and the relations of any of these changes to physical processes such as impact and accretion of mass remain uncertain.

For example, compared with the sun or undifferentiated meteorites, Earth is depleted in moderately volatile elements (for example, Rb) relative to refractory elements (for example, Sr). In undifferentiated materials, the Rb/Sr ratio is relatively high and so too is the rate of growth of the $^{87}Sr/^{86}Sr$ ratio due to decay of ^{87}Rb. At some point, Earth's materials experienced a transition to the much lower present Rb/Sr ratio and thus to much slower growth of $^{87}Sr/^{86}Sr$. The initial $^{87}Sr/^{86}Sr$ ratio for a volatile-poor planet such as Earth is usually taken to reflect the time of this transition, which is commonly considered to have occurred in a dispersed state before planetary accretion but which may be argued to have occurred as a result of accretion. The same might be true for the U (refractory)–Pb (volatile) system, but there is an added complication. If most of Earth's Pb is in the core, then our Pb chronology—which relies on outer Earth materials—might instead be dating core formation. The I-Pu-Xe system may be dating preaccretionary volatile separation as Pu is refractory, I is volatile, and Xe is even more volatile. But it might alternatively record loss of atmosphere after accretion was completed, for example, in a major late impact. Hf and W are both refractory but Hf is strongly lithophilic ("rock-loving") and W is moderately siderophilic ("metal-loving"). Their separation is usually taken to reflect metal-silicate fractionation, that is, core formation. If Hf-W separation occurred early, while ^{182}Hf was still present, the silicate will display excess ^{182}W and the metal will be deficient in ^{182}W (see the figure). In contrast, if separation occurred only late, after ^{182}Hf had already decayed, silicate and metal will both have the same composition. W in Earth's mantle has the same composition as W in undifferentiated meteorites (see the figure), thus indicating late core formation, but again this must be qualified by the possibility that it was only the last few percent of the W to accrete experienced such late metal-silicate formation.

If the moon were an independent planet, we would likely conclude that it was older than Earth. Similar to Earth's estimated age, the oldest lunar rocks (*9*), likely representing the first rocks formed from a lunar magma ocean, are about 100 million years younger than the solar system. But the moon's initial $^{87}Sr/^{86}Sr$ ratio is very low (*10*), indicating that it was depleted in volatiles within a few million years of the formation of the solar system. Also, in contrast to Earth, some of the moon's surface rocks do contain excess ^{182}W (see the figure), indicating preservation of reservoirs since the time when ^{182}Hf was still extant and could generate excess ^{182}Hf after metal-silicate fractionation. If the moon was indeed formed by a giant impact when the still-accreting Earth was struck by a planetary body nearly half its size (*11*), it seems most likely that this impact occurred about 100 million years after the solar system formed, that the moon was made largely from the impactor, and that the impactor had previously enjoyed an existence as an independent planetary body for a considerable time.

To help tie down the details, we need not so much more isotopic data but a better understanding of how the physical events that define planetary histories affect the chemical events and the radionuclide systems by which we reconstruct geological time. For testing the giant impact scenario (*1*) in particular, it would be useful to have a quantitative theory for whether a preexisting atmosphere is lost in the impact, whether preexisting planetary structures (core, mantle, and crust) are reequilibrated after such an impact, and how much of the moon comes from the impactor and how much comes from the target.

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PERSPECTIVES: CONDENSED MATTER PHYSICS

Pumping Electrons

B. L. Altshuler and L. I. Glazman

We usually associate electric current with a dissipative motion of electrons (*1*): The energy provided by an external source eventually becomes heat. However, examples of a current flowing without dissipation are also well known. For instance, a static magnetic field magnetizes a metal specimen by causing edge electric currents (Landau diamagnetism) (*1*). Moreover, a static magnetic flux threading a metallic ring induces a persistent current (*2*). The electrons remain in equilibrium, and energy does not dissipate. Pumping is also a way of transferring electric charges, but it is qualitatively different from both mentioned mechanisms. An experimental investigation of electron pumping through a quantum dot is reported by Switkes *et al.* (*3*) on page 1905 of this issue.

B. L. Altshuler is in the Physics Department, Princeton University, Princeton, NJ 08544, USA, and at the NEC Research Institute, Princeton, NJ 08540, USA. L. I. Glazman is in the Physics Department, University of Minnesota, Minneapolis, MN 55455, USA. E-mail: glazman@physics.spa.umn.edu

In pumping, periodic (ac) perturbations of the system yield a dc current. This current, though not an equilibrium response to an external perturbation, may be entirely adiabatic: The system always remains in the ground state. In contrast to the more familiar dissipative rectification of ac current, the charge transferred in each cycle of adiabatic pumping is independent of the period T. Therefore, at large T, pumping dominates.

A remarkably clear theoretical example of adiabatic electron pumping was given by Thouless (*4*). For simplicity, consider spinless electrons in a one-dimensional channel, subject to a potential $U(x)$ periodic along the channel: $U(x + a) = U(x)$. Provided the number of electrons na per period a equals an integer N, the lowest N bands of the energy spectrum are full while the higher bands are empty. Now let the potential move with some small velocity, $U(x,t) = U(x - vt)$. At each point x the potential $U(x,t)$ varies periodically with time, because $U(x)$ is translationally symmetric. If elec-