The articles in this Elements issue highlight various aspects of the formation and evolution of the Earth’s earliest crust. This Toolkit includes a glossary of terms, concepts, and analytical approaches important to the study of Archean cratons to provide context for the methodologies discussed in these articles.

**TIMESCALE OF EARLY EARTH**

Geologic timescales are rarely presented without compressing the Precambrian portion of Earth history. For this reason, it is easy to overlook the fact that the time between Earth formation and the end of the Archean at 2.5 Ga (billion years ago)—the time period covered by the articles in this issue—represents fully 45% of Earth history (Fig. 1).

The geologic timescale for the earliest part of Earth history is divided into two eons: the Hadean (4.6 to 4.0 Ga) and the Archean (4.0 to 2.5 Ga) (Fig. 1). The Hadean is defined as the time from Earth formation to the earliest part of the rock record. We use the 4.568 billion-year-old age of the oldest grains condensed from the solar nebula found in meteorites (Ca-Al-rich grains included in chondritic meteorites) to mark the start of Earth accretion (Bouvier and Wadhwa 2010). The beginning of the Archean has been established at 4.031 Ga based on the 10 oldest U-Pb zircon ages from the Acasta gneiss complex in the Slave craton of northern Canada (Bowring and Williams 1999). Consequently, we refer to all ages older than this as Hadean.

The Archean eon is divided into two eras: the Neoarchean (2.8–2.5 Ga) and the Palearchean (4.0–2.8 Ga). The Palearchean is further divided into the early (3.8–2.8 Ga) and late (2.8–2.5 Ga) periods. The Neoarchean era is divided into the early (2.8–2.7 Ga) and late (2.7–2.5 Ga) periods. The Phanerozoic eon is divided into the Paleozoic (541–252 million years ago), Mesozoic (252–66 million years ago), and Cenozoic (66 million years ago to the present). The Phanerozoic eon is further divided into five periods: Cambrian (541–485 million years ago), Ordovician (485–443 million years ago), Silurian (443–416 million years ago), Devonian (416–354 million years ago), and Carboniferous (354–299 million years ago).

**Earth’s Composition and Reservoirs**

It is not possible to determine the average geochemical composition of the entire Earth directly, so its overall composition—often expressed as the bulk Earth composition—is estimated from the composition of Type-I carbonaceous chondrite meteorites, a class of meteorites that is thought to have formed during condensation of the solar nebula and formation of the Solar System and has not undergone melting or metamorphism since that time. This composition is also referred to as the chondritic uniform reservoir (CHUR).

Very early in its history, the metallic core separated from the silicate portion of the Earth (Nimmo and Kleine 2015). Thus, it is useful to define the geochemical composition of the bulk silicate Earth as the composition of the entire Earth minus the core. The composition of the bulk silicate Earth is equivalent to that of the primitive mantle.

Today, the Earth is composed of a number of geochemically distinct components, including the core, mantle, and crust. These are commonly referred to as reservoirs. The portion of the mantle that partially melted to produce magmas that contributed to forming the Earth’s crust is known as the depleted mantle. The time when the depleted mantle reservoir began to form and the evolution of its composition and volume over Earth history are intricately related to the formation and evolution of the complementary crustal reservoir (Fig. 2). These processes are incompletely resolved in time and space, as discussed by O’Neil et al. (2024 this issue) and Rey et al. (2024 this issue).

**Figure 2** Crustal growth models. Estimates of the time at which the crust first separated from the mantle vary greatly, as does the volume of crust present throughout Earth history, which on this figure is expressed as a percentage of present-day crustal volume. This figure illustrates a variety of models ranging from those that envision abundant crust formation early in Earth history (i.e., green curve) to many other models that suggest that the volume of continental crust has increased over time, albeit at different rates. Modified from Hawkesworth et al. (2019).

**Figure 3** Schematic illustration of continental freeboard (Korenaga et al. 2017). If sea level rises by height $h$, more of the continental shelf will be submerged, the area of exposed landmasses decreases by area $A_{ex}$, and continental freeboard decreases.
The Earth’s oceans constitute another reservoir that appears to have formed early, likely by 3.7–3.8 billion years ago when pillow basalts are first preserved in the rock record (Nutman et al. 2013). The concept of crustal growth is intimately linked to the concept of continental freeboard, which refers to the mean height of the continental landmasses with respect to sea level at any given time (Fig. 3; Korenaga et al. 2017). This concept is important because freeboard reflects a balance between the volume of the oceans, the shape of ocean basins, and the volume, density, and thickness of the oceanic and continental crust. See Rey et al. (2024 this issue) for a broader discussion of changes in sea level during the Archean.

GEODYNAMICS OF THE EARLY, NON-UNIFORMITARIAN EARTH

Geologic processes are commonly assumed to be uniformitarian, that is, we can infer how ancient rocks were formed and modified by looking at modern processes, including plate tectonics, that affect the Earth today. This assumption cannot be applied to the first 1.5 to 2 billion years of Earth history. For example, the formation of Earth’s core, silicate magma oceans, and the Moon were early, one-time events. Exponential, secular cooling of the Earth as radioactive heat production decreased produced different temperature regimes at different times in the past (Korenaga 2013). This cooling also likely affected Earth’s geodynamics, a general term that refers to the physical stresses that deform Earth’s surface and interior. Instead of stiff plates of crust moving horizontally above a convecting mantle in a modern so-called mobile lid process, Hadean and Archean tectonics may have been driven by mantle plumes beneath a weaker, stationary crust, in a so-called stagnant-lid scenario (Fig. 4A). Instead of subduction, the early crust may have been disrupted and even relocated in response to mantle upwellings and downwellings that have been linked to processes such as sagduction and drip tectonics (Fig. 4B; Martin and Arndt 2021).

In addition, studies of the lunar surface indicate by analogy that impacts of large, extraterrestrial bodies (e.g., planetesimals and large meteorites) strongly affected the early Earth, but the rate of these impacts tailed off by the end of the Archean (see recent compilation on Fig. 2 in Tai Udovicic et al. 2023). Large impacts on the early Earth, including the Moon-forming impact, may have generated enough heat to wholly or partially melt large parts of the mantle. The resulting “magma oceans” may have persisted for millions of years and greatly influenced the chemical and physical evolution of the mantle.

The time at which modern plate tectonics initiated is a subject of considerable debate (Korenaga 2013). As Rey et al. (2024 this issue) point out, it is possible that there was a period when both stagnant and mobile tectonic processes were operating in parallel, a situation they refer to as dual-mode geodynamics. The transition from stagnant to mobile lid tectonics need not have been synchronous globally, but likely occurred at different times in different parts of the Earth and affected individual cratons differently (Frost and Mueller 2024 this issue).

AGE DETERMINATIONS OF ARCHEAN AND HADEAN EVENTS

U-Pb in Zircon

The U-Pb system in zircon has evolved to become the most reliable method for obtaining accurate and precise ages of events in the geologic past, particularly the Precambrian. The two radioactive isotopes of uranium (235U and 238U) decay at vastly different rates, providing two independent age determinations and an opportunity to compare ages from one decay system to the other in a unique cross-check. When the ages agree within analytical errors, the ages are considered concordant, implying that the zircon has behaved as a closed system since it formed. The data are often depicted on concordia diagrams, on which curved lines give the ages of concordant data regardless of the ratios used to define concordia. Two different common concordia plots are shown in Figure 5. Discordant ages (ages from the two decay systems that do not agree) are also valuable because the lack of agreement often can be related to loss of Pb from the system (e.g., a zircon grain). Straight lines through discordant points (discordia; Fig. 5) on the concordia plots may intersect concordia at two times, and can then identify both the time of original crystallization and the time of the event that caused Pb-loss, such as metamorphism. Because ancient zircons likely experienced more than one Pb-loss event, the lower intercept of the discordia line with concordia may have only limited or no geologic meaning.

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**Figure 4** Earth currently dissipates >90% of its internal heat via a “multi-plate mobile lid” system (plate tectonics) driven by the sinking of cold dense lithosphere into the convecting mantle. In the Archean-Hadean, higher mantle temperatures and more rapid convective heat loss may have occurred in a “stagnant lid” system, which transitioned to plate tectonics over time. Two possible transitional stages include: (A) Hadean mafic crust with nascent TTG differentiates (yellow) formed over broad mantle upwellings (orange arrows) and (B) An Archean stage when dense volcanic rocks of greenstone belts sank through warm, weak felsic crust, displacing older felsic crust as diapirs with melt (purple) that intruded the founding greenstones. This process, often labeled “sagduction,” produced the dome and keel pattern shown in Figure 2 of Frost and Mueller (2024 this issue).

**Figure 5** Two commonly used types of concordia diagrams. (A) Tera-Wasserburg concordia diagram and (B) Wetherill concordia diagram. In each, the concordia is shown as the blue curve. Samples plotting on the concordia (green ellipses) give the same age in both the $^{238}$U-$^{206}$Pb and $^{235}$U-$^{207}$Pb decay systems. Arrays of discordant points (open purple ellipses) may form a chord (discordia) that intersects the concordia at the crystallization age. See Figure 4 in Laurent et al. (2024 this issue) for an example of data interpreted on a Tera-Wasserburg diagram.
Common Pb and the Age of the Earth

Common Pb isotopes are those found in “common” rocks and minerals that have not necessarily remained closed systems since they formed. The Pb isotopes produced by decay of 235U (207Pb) and 238U (206Pb) are typically reported as ratios relative to the non-radiogenic isotope 204Pb. Lead produced from decay of 232Th is similarly reported as 205Pb/204Pb from the single decay of 232Th to 208Pb. The Pb isotopic composition of an individual rock or mineral evolves with time to higher ratios of radiogenic Pb (206Pb, 207Pb, 208Pb) to non-radiogenic 204Pb along curved paths called growth curves (Fig. 6). The relatively rapid decay of 235U means that ~80% of all 235U on Earth had decayed by the time the Earth formed and is the reason why the growth curves shown on Figure 6 flatten out at younger ages. Consequently, rocks and minerals with high 207Pb/204Pb relative to 206Pb/204Pb indicate interaction with high-U/Pb reservoirs (that must have formed in the Hadean or Archean). These are referred to as high-μ or early enriched reservoirs. These reservoirs likely formed by magmatic differentiation within the crust because U is preferentially partitioned into magmas to a greater extent than Pb in crustal and mantle melts.

Common Pb isotopes have played an important role in our evolving understanding of Earth and Solar System history. For example, Pb isotopic data from galenas, oceanic sediments, and meteorites were combined to provide the first viable determination of the age of the Earth (Fig. 6B; Patterson 1956).

**Figure 6**

(A) Evolution of Pb isotopes in Earth reservoirs with different U/Pb ratios (μ = 235U/204Pb) over Earth history from an initial bulk Earth Pb isotopic composition estimated from iron meteorites (Blichert-Toft et al. 2010). At any given time, rocks from reservoirs of the same age, but with different μ-values, will define a straight line (an isochron) with a slope that is a function of age. (B) The slope of the red line defined by the Pb isotopic composition of meteorites (blue symbols; blue square = iron meteorites; blue circles = stony meteorites) gives the time of Solar System formation. Because average deep-sea sediment, taken as a proxy for average continental crust, and recently formed galena samples (μ = zero) from ore deposits lie on the same isochron as meteorites, the Earth and meteorites were interpreted to have the same age (Patterson 1956).

**ISOTOPIC SYSTEMS USED TO INFERENCE EARTH DIFFERENTIATION PROCESSES**

In addition to the U-Pb system, many other radiogenic parent-daughter isotope pairs are used to determine both the timing and nature of events on Earth. Most of these involve parent isotopes with slow decay rates, such as 147Sm-143Nd (half-life (T½) of 147Sm = 107.0 billion years), 176Lu-176Hf (T½ of 76Lu = 37.2 billion years), and 187Re-187Os (T½ of 87Re = 41.6 billion years). As with Pb isotopes, the daughter isotope is reported relative to a stable isotope of the same element. Examples include 143Nd/144Nd (shown on Fig. 7A), 176Hf/177Hf, and 187Os/188Os. The variation in these ratios for Earth materials is very small, so it has become customary to report measured values relative to a standard such as CHUR (the chondritic uniform reservoir, which approximates the bulk silicate Earth composition). This is called the epsilon notation, and is given by the formula:

\[
\epsilon_{\text{Nd}}(t) = \frac{^{143}\text{Nd}_{\text{sample}}(t)}{^{144}\text{Nd}_{\text{CHUR}}} - 1 \times 10,000
\]

where \( t \) is the time of interest. Multiplying by 10⁴ results in values for \( \epsilon_{\text{Nd}}(t) \) commonly between +15 (depleted mantle) and −50, (old felsic crust), where positive values denote samples with higher 143Nd/144Nd than the bulk Earth, and negative values identify samples with lower 143Nd/144Nd than the bulk Earth (Fig. 7B). This is the case because mantle melts have lower Sm/Nd than the residual (depleted) mantle. Analogous expressions give \( \epsilon_{\text{Hf}}(t) \). In the 187Re-188Os system, the \( \epsilon_{\text{Os}}(t) \) notation refers to percent deviation from CHUR of 187Os/188Os (multiplying by 100 instead of 10,000).

When evaluating diagrams and/or text that use the epsilon notation, it is important to remember that the relative rate of change of a sample compared to a model reservoir or another sample is related to the half-life of the parent isotope, the time since mineral/rock formation, the parent/daughter ratio, and the assumptions used to calculate the \( \epsilon \) of model reservoirs. For example, the depleted mantle formed at 4.6 Ga with an epsilon of zero and has evolved to a range of values centered about +13 εNd. Evolution diagrams may be plotted with measured isotope ratios (Figs. 7A) or derivative ratios such as εpsilons (Figs. 7B). Although Hf and Nd isotopic compositions of rocks and minerals can be measured very precisely today, extrapolating these values back over 3–4 billion years involves uncertainties beyond analytical errors.

**Model Ages**

As shown in Figure 7B, a model age is simply an estimate of the time that a sample was extracted from a model reservoir. For example, a given sample’s measured 143Nd/144Nd is extrapolated back in time using...
the sample’s measured $^{147}\text{Sm}/^{144}\text{Nd}$ until its Nd isotopic composition matches that of the evolving reservoir, commonly the depleted mantle. That time is referred to as the depleted mantle model age: $T_{\text{DM}}$. A graphic representation of this concept is shown as point $T_{\text{DM}}$ on Figure 7B, where the model mantle evolution curve is intersected by the evolution curve of an individual sample at time $T_2$.

**Extinct Radionuclides**

Some radioactive isotopes produced during stellar processes have such short half-lives that they have completely decayed and are no longer present in our Solar System. These are known as extinct radionuclides. Extinct radionuclides are especially well suited for understanding events that took place on the early Earth while these short-lived radioactive isotopes were still “alive.” For example, consider the decay of lithophile (crust-loving) $^{182}\text{Hf}$ ($T_{1/2} = 8.9$ million years) to siderophile (iron-loving) $^{182}\text{W}$. In this system, $W$ is fractionated into the core and Hf into the mantle during core formation. The fact that $^{182}\text{W}/^{184}\text{W}$ is not completely homogenized in the modern mantle after 4.5 Ga of convection and is more heterogeneous in Archean rocks than modern mantle-derived rocks help us understand the timing of core formation and the extent of mantle mixing over time (Fig. 8). Another extinct radionuclide, $^{146}\text{Sm}$, has also been applied to the chronology of early and Hadean to Archean mantle-derived rocks.

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**Oxygen Isotopes**

The most abundant element in the silicate Earth is oxygen. The isotopes of oxygen are all stable, yet their abundance varies in Earth materials because the mass difference between isotopes results in different equilibrium and kinetic behavior. A delta notation ($\delta$) is used for oxygen isotope ratios, which most commonly compare the $^{18}O/^{16}O$ of a sample to a standard material, typically “standard mean ocean water” (SMOW), and is expressed as per mil deviations ($10^3$) from the standard’s ratio. Samples with positive $\delta^{18}O$ have higher $^{18}O/^{16}O$ than the standard, and samples with negative $\delta^{18}O$ have lower $^{18}O/^{16}O$ than the standard. In zircon, $\delta^{18}O$ higher than values for the depleted mantle (5.3‰ ± 0.6‰) can indicate incorporation of a sedimentary component and/or interaction of their parent magmas with meteoric water. Figure 9 shows a compilation of oxygen isotope data for zircons of different ages depicting a distinct change across the Archean-Proterozoic boundary.

**Presentation of Trace Element Geochemical Data**

A common approach to constraining the origin of (meta)igneous and (meta)sedimentary rocks is through the use of trace elements, i.e., elements present below 0.1% by weight. These abundances are often reported in parts per million (ppm) and shown graphically as ratios relative to chondritic meteorites and major Earth reservoirs, such as estimates of the primitive mantle, average crust, average shale, etc. (Fig. 10). The regular behavior of the rare earth elements (REE) is particularly useful because the contraction of ionic size with increasing atomic number leads to predictable behavior in many systems, from igneous to sedimentary. Be aware that the normalizing composition grossly affects the shape of the patterns and that the y-axis is always logarithmic. Moreover, the REE are often discussed under the broader term “immobile elements,” which refers to elements that do not readily substitute for the major cations in rock-forming minerals. Depending on a rock’s mineralogy, an element may be incompatible in one rock, but not in another.
lithospheric-scale structures, tomographic methods are dependent on the number and distribution of seismometers or seismic stations that record the observations. The data recorded on these instruments, such as earthquake waveforms, or even “noise” generated by the oceans, can be used to illuminate the changes in seismic velocities of materials in the subsurface. Body wave tomography, which uses the arrival times of P or S waves, and surface wave tomography, which uses the dispersive nature of Rayleigh or Love waves, have quite different resolutions and sensitivities. The resolution of the models generated is primarily controlled by the wavelength of the seismic waves used in the tomographic inversion, regardless of the specific technique or seismic wave type analyzed. Seismic waves are only sensitive to, or able to detect, heterogeneities that have length scales that are similar to the seismic waves themselves. For example, a 1-second (1-Hz) P-wave with a velocity of 5 km/s has a wavelength of 5 km, which roughly corresponds to the size of the smallest detectable heterogeneity. Surface wave tomography can produce cross sections similar to those shown in Figure 3 of Cooper and Miller (2024 this issue). These cross sections are based on much longer wavelength Rayleigh waves, providing lateral resolution on the scale of 10s of kilometers. Because of the way these waves propagate through the Earth, they are most sensitive to lateral variations in the lithosphere and less sensitive to sub-horizontal structures, such as the crust-mantle boundary (Moho). Other types of seismic imaging techniques that use body waves (P and S waves), which travel along more vertical paths, are more sensitive to boundary layers. These methods, typically referred to as receiver functions, are more often used for determining the depth to the Moho or the lithosphere-asthenosphere boundary. Regardless of methodology, these seismic images can then be tested against the temperature and pressure estimates provided by crustal and mantle xenoliths.

**GEOPHYSICAL IMAGING OF THE EARTH’S STRUCTURE**

**Seismic tomography** is not only a method to image the Earth’s interior, but also a term that refers to a collection of different modeling methodologies centered around an inverse problem. Solutions to inverse problems involve using observations to create a model of the conditions that created the observations. For imaging crustal- or