


**Cross-references**

Albedo Feedbacks  
Antarctic Glaciation History  
Arctic Sea Ice  
CLIMAP  
Cryosphere  
Diatoms  
Last Glacial Maximum  
Marine Biogenic Sediments  
Paleoclimate Modeling, Quaternary  
Radiolaria  
Transfer Functions

**ARCHEAN ENVIRONMENTS**

A primary research objective in the study of Archean rocks (>2.5 Ga) has been to determine conditions on the surface of the Earth during the first two billion years of its history. Recent advances in isotope geochemistry and geochemical microanaly- sis have allowed new insights into the earliest few hundred million years of Earth’s history, and have pushed back the time at which the first life could have emerged.

The earliest subdivision of the Archean Eon beginning with Earth’s accretion at 4.56 Ga and extending to ca. 4.0–3.8 Ga, has been called both the Priscoan (“previous time”) Era and Hadean (“Hell-like”) Era. These divisions of geologic time are based on the inherent bias of the rock record, making distinctions in part based on whether or not rocks have survived. The term “Hadean” is also based on the preconception that Earth was so hot from formation and bombardment by meteorites that the atmosphere was dominated by steam and rock deb- ris, and solid material would have been pulverized or melted (Cloud, 1972). Detrital zircon crystals (ZrSiO4) from this period challenge this preconception (Compston and Pidgeon, 1986; Mojzsis et al., 2001; Peck et al., 2001; Wilde et al., 2001; Valley et al., 2002; Cavosie et al., 2004, 2005), and provide evidence for proto-continental crust and low temperatures at the surface of the Earth during the period 4.4–4.0 Ga.

**4.57 to 4.45 Ga: the hot early earth**

The most widely-cited age of the Earth is ca. 4.55 Ga as deter- mined by Claire Patterson from lead isotopes in primitive meteorites and terrestrial sediments (Patterson, 1956). This is the age of condensation of solid material from the solar nebula (since revised to 4.56–4.57 Ga), and dates the origin of the material from which Earth began to accrete.

There are no recognized terrestrial samples from the first 150 Myr of Earth’s history, thus constraints on surface condi- tions are inferred from isotopic data and the geologic history of the Moon. The currently accepted model of Moon formation calls for impact of the Earth with a planet-sized bolide, causing separation of the Moon from Earth’s mantle (Canup and Righter, 2000). This model explains the orbit of the Moon and its geochemical similarity to Earth. The oldest Moon rocks have ages ca. 4.49 Ga, constraining the age of the catastrophic impact (Canup and Righter, 2000) and in agreement with W isotopes, which suggest formation of the Moon within 30 Myr of the Earth’s accretion (Halliday, 2000).

Heat from meteorite impacts, residual heat from the Earth’s accretion and core formation, and high heat production from radioactive elements likely made the surface unsuitable for the formation of oceans and continents during the beginning of Earth’s history. For example, estimates of heat production from the early Earth’s store of radioactive elements are on the order of 3–6 times that of today (Pollock, 1997). Conditions on the surface may have been extreme enough to form magma oceans until the waning of large impacts and the dissipation of heat into space (cf. Jones and Palme, 2000). Dissipation of heat could have resulted in rapid and turbulent mantle convection well into the Archean (Pollock, 1997). The last large terrestrial impact could have caused the resetting (and presumably planet-scale mixing) of xenon and other isotope systems at 4.45 ± 0.05 Ga (Zhang, 2002). It is possible that a slightly older impact formed the Moon. No terrestrial materials appear to have survived this last catastrophic event.

**The cool early earth: evidence from 4.40 to 4.00 Ga detrital zircons**

In the early 1980s newly developed ion microprobe technology was used to identify individual zircon crystals older than 4.0 Ga (Froude et al., 1983; Compston and Pidgeon, 1986). These crystals were eroded from their original host rocks, trans- ported by water ca. 3 Ga, and are now found in metasedimentary rocks of the Narrrary Gneiss Complex (Western Australia). The orig- inal ≥4.0 Ga igneous rocks are not known to have survived ero- sion and recycling. These Zircon crystals provide the only direct evidence for conditions at the surface of the Earth ≥4.0 Ga, as no other samples of solid material are known from this time period.
Recently, a crystal dated at 4.4 Ga was discovered (Wilde et al., 2001), providing a glimpse of conditions within 160 Myr of the Earth’s accretion.

Zircons from Western Australia are evidence for 4.4–4.0 Ga rocks similar to modern continental crust. Mineral inclusions in the 4.4 Ga zircon include SiO₂ (Peck et al., 2001; Wilde et al., 2001; Cavosie et al., 2004), indicating crystallization from an evolved, silica-saturated magma. Similar conditions have been inferred for other Archean magmas based on inclusions from other 4.2–4.0 Ga zircons (Maas et al., 1992). Trace element compositions (including rare earth elements) of these zircons are also elevated relative to the mantle, consistent with crystallization from an evolved magma (Maas et al., 1992; Peck et al., 2001). All of these lines of evidence suggest the existence of igneous rocks similar to those of modern continental crust.

Isotopic analysis of these 4.4–4.0 Ga crystals has provided new constraints on surface conditions of the Earth (Peck et al., 2000, 2001; Mozsis et al., 2001; Wilde et al., 2001; Cavosie et al., 2005). Oxygen isotope ratios (³¹⁸O/³²⁰O) of these crystals are elevated relative to the primitive ratios of the Earth’s mantle and the Moon (Valley, 2003). Elevated oxygen isotope ratios in terrestrial materials are most typically caused by low-temperature interaction between rocks and water at the surface of the Earth (i.e., hydrothermal alteration or low-temperature mineral formation). Magmas with elevated oxygen isotope ratios are formed by melting or assimilation of this material. The possible presence of liquid water calls into question some widely assumed conditions of the “hell-like” nature of the early Earth (Valley et al., 2002).

**Emergence of continental crust**

The first igneous rocks on Earth were most likely basaltic or komatiitic, with similarities to modern oceanic crust (e.g., Taylor, 1992). Lack of preservation of this crust may be due to melting and recycling back into the mantle. The timing and development of continental crust is a fundamental question as to surface conditions of the early Earth. The oldest recognized continental crust is 4.03 Ga granitic gneiss preserved in the Acasta Gneiss Complex, Northwest Territories, Canada. These strongly deformed rocks constitute <20 km² of the 4.0–2.6 Ga Slave Craton. Other remnants of early (≥3.8 Ga) crust are limited to a few thousand square kilometers in the Ulvaq gneisses of Labrador, the Qianxi and Anshan Complexes in China, the Napier Complex of Antarctica, and the Isua Gneiss Complex and Aasivik Terrane of southwestern Greenland (Figure A23). Sedimentary rocks ca. 3.5–3.0 Ga in age contain some of the oldest recognized terrestrial material: 4.4–3.2 Ga detrital zircons in the Narryer Gneiss Complex, detrital zircons as old as 4.0 Ga from the Wyoming Province, USA, and detrital zircons as old as 3.85 Ga from metasediments in the Qianxi complex (see Nutman et al., 2001). The Archean rock record is more complete for rocks ca. 3.5 Ga and younger.

The small amount of early Archean crust can either be interpreted as the surviving pieces of once extensive continental crust or as a reflection of the low rate of continental crust production in the early Archean. This dichotomy is reflected in different crustal growth models developed over the last 30 years: some propose rapid generation of Archean crust (most of which would be recycled), while others call for initially slow growth. Studies of middle Archean sediments support this latter view (e.g., Nutman et al., 2001), as recycled early Archean rocks constitute a minor component. This contrasts with geochemical signatures from the Acasta Gneiss, the Isua Gneiss Complex, and ≥4.0 Ga detrital zircons that show derivation from a mantle which had already produced crust earlier in its history (≥4.3 Ga; see Jacobsen, 2003 and references therein). Taken with the continental affinities of the Jack Hills detrital zircons, there is now ample evidence for continental crust between 4.4 and 4.0 Ga. Most of this original crust was eroded and/or recycled back into the mantle, perhaps aided by meteorite bombardment of the early Earth.

**Timing of meteorite bombardment**

Meteorite bombardment of the Earth during the period 4.4–3.8 Ga is known from the cratering record of other bodies in the inner solar system, but the terrestrial record has been erased by plate tectonics and erosion. On the Moon (where impact events have been dated) the time-integrated flux of impacts is estimated through crater density, addition of material to lunar crust, and the extent of impact stirring (Arrhenius and Lepland, 2000; Hartmann et al., 2000).

The terrestrial meteorite flux over time is controversial and has not been uniquely determined. Presumably meteorite bombardment would have been at its peak early in the Solar System’s history, during and preceding extraction of the moon at ca. 4.5 Ga. The waning of impacts is not well constrained, although many lunar samples of impact glass yield a grouping of ∼3.9 Ga ages called the “late heavy bombardment” (e.g., Cohen et al., 2000). If the rate of impacts has decreased steadily through time, then the high rate of impacting at 3.9 Ga suggests even more impacts earlier.

However, recent dating of lunar impact glasses supports an alternate hypothesis: that the cluster of 3.9 Ga ages represents a renewal of impact intensity caused by deflection of materials from the outer Solar System (Cohen et al., 2000). If the Earth was subjected to a spike of meteorite bombardment ca. 3.9 Ga, then the 4.4–4.0 Ga interval might have been relatively tranquil, consistent with cooler surface temperatures inferred from zircons crystallized during this period (see Valley et al., 2002). The terrestrial record of bombardment is equivocal. There are no reported descriptions of shock features in 4.4–3.8 Ga zircons or rocks, as would be expected if they had been involved in large impacts. Metasedimentary rocks from southwestern Greenland (3.8–3.7 Ga) do not contain impact-related sedimentary structures (i.e., surge deposits, see Arrhenius and Lepland, 2000), nor has measurable extraterrestrial iridium...
been found. However, evidence has been presented for an elevated extraterrestrial tungsten isotope signature in Isua metasediments of Greenland (Schoenberg et al., 2002), similar to evidence for extraterrestrial chromium isotopes associated with ~3.2 Ga South African beds of impact glass (Kyte et al., 2003).

The timing of the early bombardment of Earth is poorly constrained and limited by the rocks available for study. However, indirect geologic and geochemical evidence for liquid water during this period argues for at least transient periods of quiescence during bombardment. It is possible that transient periods allowed the precipitation of oceans between ~4.4 (the first evidence for continental crust) and 3.9 Ga (the late heavy bombardment).

**Oceans**

A critical parameter for evaluating surface conditions on Earth is whether or not liquid water is a stable phase, a function of both temperature and vapor pressure of the atmosphere. Supracrustal rocks in the Itsaq Gneiss Complex (the Isua and Akilia Island localities) are the earliest known rocks deposited in water: 3.8–3.7 Ga pillow lavas and chemical sediments (Myers and Crowley, 2000). The age and ultimate origin of rocks of the Akilia Island locality are controversial, but the Isua lithologies are widely accepted as having formed under submarine conditions (below wave base according to Fedo et al., 2001). The next oldest unequivocal evidence for submarine deposition, although indirect, is the abundance of metasediments ≤3.5 Ga worldwide. Volcanogenic massive sulfide deposits are also common in rocks ≤3.4 Ga and could have formed at a variety of ocean depths (including deep water ≥1,000 m; Rasmussen, 2000).

There are critical gaps in the rock record from ≥3.8 Ga and between 3.7 and 3.5 Ga, a period during which life is believed to have originated. A proxy for the presence of water is the geochemical signature of water-rock interaction associated with Archean detrital zircons (Peck et al., 2001; Mojzsis et al., 2001; Wilde et al., 2001; Valley et al., 2002; Cavosie et al., 2005). The elevated oxygen isotope ratios in these 4.4–4.0 Ga zircons are consistent with low-temperature exchange between rocks and water. On the modern seafloor, hydrothermal alteration ≤ ca. 250 °C produces similar oxygen isotope ratios, but there is more uncertainty as to the conditions of water-rock interaction during the Archean. Lower temperature estimates are plausible if: (a) oxygen isotope ratios of the oceans have slowly risen over time (e.g., Perry, 1967), (b) meteoric waters instead of seawater were involved in Archean water-rock interaction, or (c) hydrothermal alteration was less efficient than today (Valley et al., 2002). Steam and liquid water could have coexisted at temperatures ≤250 °C, especially given the likelihood of periodic meteorite impacts. However, steam alone is an inefficient mechanism for exchange between water and rocks and a steam atmosphere the size of modern oceans would require temperatures outside of the oxygen isotope constraints from the detrital zircons (Valley et al., 2002).

The oxygen isotope record from zircons records the compositions of magmas from 4.4 Ga to the present (Figure A24). This record includes 4.4–4.0 and 3.6–3.2 Ga detrital zircons, and zircons in igneous rocks from ≤3.6 Ga. The Archean zircons came from igneous rocks that had oxygen isotope ratios ranging from mantle-like to elevated values, similar to Phanerozoic rocks but with a more limited spread (Valley, 2003). These similarities would not be expected if older rocks had formed in the absence of water and only younger ones had interacted with water (see also Campbell and Taylor, 1983).

The constancy of isotope ratios suggests that surface temperatures from 4.4 to 4.0 Ga were similar to those of the later Archean. The increase in values starting at the Archean-Proterozoic boundary marks an increased involvement of mature supracrustal rocks (with a water-rock interaction signature) in the genesis of igneous rocks, but it is not certain if this indicates increased rates of recycling or more highly evolved compositions (Peck et al., 2000).

The temperature of Archean oceans has been controversial because 3.8–2.5 Ga chemical sediments have low oxygen isotope ratios compared to younger rocks (see Perry, 1967; Perry and Lefticaru, 2003). This has been taken to indicate any (and all combinations) of the following factors: (a) Archean ocean water temperatures of up to 80 °C, (b) lower oxygen isotope ratios of Archean ocean water, and (c) later alteration of original isotope ratios in ancient samples. The oxygen isotope ratio of the ocean seems to have been buffered at or near present levels since at least 3.6 Ga, as predicted by models and observed in hydrothermally altered rocks (e.g., Muehlenbachs, 1998). In addition, careful sampling in ~3.5 Ga rocks has constrained the influence of post-depositional alteration. Although the oxygen isotope ratio of the ocean has certainly fluctuated within a relatively small range and some chemical sediments may have been altered, the overall evidence points towards ocean temperatures up to ca. 40 °C higher than those today as late as ~3.5 Ga (Knauth and Lowe, 2003; Perry and Lefticaru, 2003).

What was the thermal history of the hydrosphere during the Archean? During the period of accretion and heaviest bombardment of the Earth (≥4.5 Ga) all water must have been vapor or dissolved in silicate melts. Oxygen isotope evidence from 4.2 to 4.0 Ga zircons indicates a liquid or liquid-steam hydrosphere ≤250 °C. This is consistent with predictions of rapid cooling of the surface of the Earth (1–10 Myr) shortly after accretion and during waning bombardment (Pollack, 1997; Zahnle et al., 1988; Sleep et al., 2001). Late heavy bombardment of Earth at ~3.9 Ga could have vaporized Earth’s oceans, but beginning at 3.6–3.8 Ga there is good geological evidence for a stable hydrosphere. It has been suggested that in order for oceans to remain liquid rather than freezing, the lower luminosity of the Sun during the Archean may have been offset by high levels.

![Figure A24](image-url)
of greenhouse gases (Sagan and Chyba, 1997), or that oceans may have been partially covered by ice and warmed by submarine volcanic and periodic impacts (Bada et al., 1994).

**Evidence and controversy for early life**

The need to understand the timing of and conditions for the origin of life is one of the principle motivations for assessing the early Archean environment. The primary requirements for evolution of life are organic molecules, a source of energy, and liquid water. Organic molecules could have been delivered to Earth by carbonaceous meteorites and comets, and energy was available from the Sun and terrestrial volcanism. The precipitation of liquid water from an early steam-rich atmosphere was thus the final ingredient for an environment where life could have evolved during the period 4.4–4.0 Ga, perhaps to be extinguished by large bolide impacts or during the late heavy bombardment at 3.9 Ga. Theoretical limits of the habitability of the early Earth are thoroughly discussed elsewhere (see Nisbet and Sleep, 2001), but here we focus on evidence for early life from the rock record.

The earliest evidence for life is a distinctive carbon isotope signature from metamorphosed sedimentary rocks of southwestern Greenland. Low carbon isotope ratios (13C/12C) in reduced sedimentary carbon are commonly the result of isotope fractionation by biological pathways, and thus are taken as a “fingerprint” of ancient life even in the absence of preserved fossils (Schidlowski, 2001). While some studies have been controversial, the best evidence for early life comes from southwestern Greenland: low carbon isotope ratios of graphite particles in clastic sedimentary rocks from 3.8 to 3.7 Ga rocks at Isua (Rosing, 1999). These rocks are preserved in low-strain zones, were clearly waterlain, and the remains of various Archean microbes (Schopf, 1993). These microstructures are interpreted as being synthetic microbes. These microstructures in the Apex chert of Australia as well as numerous occurrences of stromatolites (mounds produced by photosynthetic microbes). These microstructures are interpreted as being the remains of various Archean microbes (Schopf, 1993). This claim of diversity of life has been challenged by Brasier et al. (2002), based on morphology of the features as well as a hydrothermal origin of the host-rock. However, the kergen-like Raman spectra and low carbon isotope ratios are strong evidence of biologic activity (Schopf et al., 2002). Furthermore, abundant stromatolites are found in associated rocks. The association of microfossils with seafloor hydrothermal systems has been recognized in rocks as old as 3.5 Ga (Van Kranendonk, 2001), supporting the hypothesis that life originally evolved near submarine vents or springs in early Archean oceans (e.g., Nisbet and Sleep, 2001).

William H. Peck and John W. Valley

**Bibliography**


Sagan, C., and Chyba, C.F., 1997. The early faint Sun paradox: Organic Sea ice is an extremely important component of the climate


Schoenberg, R., Kamber, B.S., Collerson, K.D., and Moorbath, S., 2002. 387 to 15 million km2 from summer to winter (see Figure A26). Sea ice is a general term that comprises several types of ice originating from the freezing of marine waters, which occurs at a water temperature of about –1.8 °C, depending upon salt content. The term “sea ice” excludes icebergs, which are massive pieces of floating freshwater ice that were calved from the front of continental glaciers. The freezing process of marine water differs from that of freshwater because it involves downward migration of brines resulting from the segregation of sea salts and ice crystals composed of almost pure water. It is a relatively complex process from a thermodynamic viewpoint since ice crystal formation induces an increase in the salt content of the surrounding seawater, thus increasing the density and decreasing the freezing point of this water.

Sea ice formation

Sea ice is a general term that comprises several types of ice originating from the freezing of marine waters, which occurs at a water temperature of about –1.8 °C, depending upon salt content. The term “sea ice” excludes icebergs, which are massive pieces of floating freshwater ice that were calved from the front of continental glaciers. The freezing process of marine water differs from that of freshwater because it involves downward migration of brines resulting from the segregation of sea salts and ice crystals composed of almost pure water. It is a relatively complex process from a thermodynamic viewpoint since ice crystal formation induces an increase in the salt content of the surrounding seawater, thus increasing the density and decreasing the freezing point of this water.

Sea ice formation is a consequence of the cooling of marine water in response to low air temperatures. It also depends upon the structure of the water mass, since the vertical density gradient actually determines the depth of the mixed surface layer, which must cool before the surface can reach the freezing point (e.g., Barry et al., 1993). A shallow stratification in the upper water column and the presence of a low-salinity (thus, density) surface water layer are important parameters for the formation of sea ice. In the Arctic Ocean, a mixed layer with salinities often below 30% at the surface, overlies a generally warmer and more saline (>34.5%) water layer originating from the Atlantic. The low salinity of the mixed layer in the Arctic Ocean results from freshwater inputs, notably from Eurasian rivers (Yenesey, Ob, Lena, Kolyma; ~1,700 km³ yr⁻¹) and the Mackenzie River (~260 km³ yr⁻¹) (Carmack, 1998).

In the Arctic Ocean, a large part of the sea ice formation occurs over shallow continental shelves surrounding the basin (Figure A25), notably in the Siberian and Laptev Seas, where the shelf is particularly wide and sea surface salinity is

Cross-references

Atmospheric Evolution, Earth

Bolide Impacts and Climate

Carbon Isotopes, Stable

Dating, Radiometric Methods

Faint Young Sun Paradox

Isotope Fractionation

Oxygen Isotopes

Stable Isotope Analysis

ARCTIC SEA ICE

Introduction

Sea ice occupies a large part of the world ocean (about 7%) and covers most of the Arctic Ocean (Figure A25), where it averages about three meters in thickness. Sea ice undergoes very large seasonal variations in its areal extent in response to changes in solar insolation. In circumpolar regions of the Northern Hemisphere, its coverage rises from approximately 7 to 15 million km² from summer to winter (see Figure A26).

Sea ice is an extremely important component of the climate system because its reflectivity plays a role in the energy budget at the Earth’s surface and because it forms a barrier between the atmosphere and the ocean, thus reducing the exchange of heat, moisture and gases. Moreover, a portion of the Arctic sea ice is exported to the North Atlantic Ocean, where it becomes a source of freshwater that can interfere with deep water formation and thus with the thermohaline circulation pattern. Recent data on Arctic sea ice derived from direct observations and satellite imagery indicate large interannual variations and suggest a decreasing trend in both area and thickness during the last decades of the twentieth century. Longer-term data on a time scale ranging from 10² to 10⁶ years available from historical archives and from deep-sea sedimentary records reveal large fluctuations in the average extent of sea ice during the Quaternary, from more reduced to much more extended covers than at present, generally as a result of hemispheric-scale climate changes. For example, historical information indicates that sea ice expanded in the subpolar North Atlantic from the Late Glacial Maximum, whereas the middle Holocene warm episode seems to have been marked by reduced sea ice cover in the Arctic Ocean and circumarctic seas. The age of initial formation of the sea ice cover in the Northern Hemisphere during the late Cenozoic remains an open question, but the development of quasi-permanent pack ice in the Arctic Ocean probably occurred during the late Miocene.

Sea ice formation

Sea ice is a general term that comprises several types of ice originating from the freezing of marine waters, which occurs at a water temperature of about –1.8 °C, depending upon salt content. The term “sea ice” excludes icebergs, which are massive pieces of floating freshwater ice that were calved from the front of continental glaciers. The freezing process of marine water differs from that of freshwater because it involves downward migration of brines resulting from the segregation of sea salts and ice crystals composed of almost pure water. It is a relatively complex process from a thermodynamic viewpoint since ice crystal formation induces an increase in the salt content of the surrounding seawater, thus increasing the density and decreasing the freezing point of this water.

Sea ice formation is a consequence of the cooling of marine water in response to low air temperatures. It also depends upon the structure of the water mass, since the vertical density gradient actually determines the depth of the mixed surface layer, which must cool before the surface can reach the freezing point (e.g., Barry et al., 1993). A shallow stratification in the upper water column and the presence of a low-salinity (thus, density) surface water layer are important parameters for the formation of sea ice. In the Arctic Ocean, a mixed layer with salinities often below 30% at the surface, overlies a generally warmer and more saline (>34.5%) water layer originating from the Atlantic. The low salinity of the mixed layer in the Arctic Ocean results from freshwater inputs, notably from Eurasian rivers (Yenesey, Ob, Lena, Kolyma; ~1,700 km³ yr⁻¹) and the Mackenzie River (~260 km³ yr⁻¹) (Carmack, 1998).

In the Arctic Ocean, a large part of the sea ice formation occurs over shallow continental shelves surrounding the basin (Figure A25), notably in the Siberian and Laptev Seas, where the shelf is particularly wide and sea surface salinity is

Cross-references

Atmospheric Evolution, Earth

Bolide Impacts and Climate

Carbon Isotopes, Stable

Dating, Radiometric Methods

Faint Young Sun Paradox

Isotope Fractionation

Oxygen Isotopes

Stable Isotope Analysis