



Ediacaran seawater temperature: Evidence from inclusions of Sinian halite

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ARTICLE INFO

Article history:

Received 21 February 2010

Received in revised form 31 August 2010

Accepted 26 October 2010

Keywords:

Ediacaran

Seawater temperature

Inclusion

Halite

ABSTRACT

Seawater temperatures throughout Earth's history have been suggested to illustrate a long-term cooling trend from nearly 70 °C at ~3500 Ma to around 20 °C at ~800 Ma. The terminal Neoproterozoic prior to the "Cambrian Explosion" is a key interval in evolutionary history, as complex multicellularity appeared with the advent of the Ediacara fauna. These organisms were likely the first that required higher levels of atmospheric and dissolved marine oxygen for their sustainability. It is known that most modern macroinvertebrates are intolerant of temperatures in excess of 45 °C. Perhaps more importantly, these high seawater temperatures limit the potential of dissolved oxygen, and therefore become an integral part of this evolutionary story. Previously, our understanding of seawater temperature during the terminal Neoproterozoic comes only from ¹⁸O/¹⁶O and ³⁰Si/²⁸Si ratios ascertained from a limited number of cherts. Isotopic ratio methods for assessing seawater temperatures are inherently indirect and have a wide range of oscillation. However, maximum homogenization temperatures (Th_{max}) of primary fluid inclusions in halite provide a direct means of assessing brine temperature, and have been shown to correlate well with average maximum air temperatures. The oldest halites date to the Neoproterozoic–lower Paleozoic (~700–500 Ma), and Ediacaran representatives can be found in Sichuan Province, China, which do preserve primary fluid inclusions for analysis via cooling nucleation methods. We utilized halite samples from the Changning-2 well, correlative to the Dengying Formation (551–542 Ma), to provide a direct assessment of terminal Neoproterozoic seawater temperature. Our measurements indicate that seawater temperatures where these halites formed are highly similar to tropical Phanerozoic seawater temperature estimates. From compiled paleotemperature data, the decline in seawater temperatures over the course of the Proterozoic, accompanied by the reduction of seawater salinity with the sequestration of salt in massive halite deposits in the Neoproterozoic, allowed the ocean system to accumulate more dissolved oxygen, and potentially paved the way for the evolutionary innovation of complex multicellularity.

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1. Introduction

The Precambrian hosted several astonishing advancements in biological evolution. Indeed, this timeframe encompasses four of the six evolutionary megatrajectories in the history of life, including the evolution of the last common ancestor of extant life, the metabolic diversification of bacteria and archaea, the evolution of the eukaryotic cell, and the origin of multicellularity (Knoll and Bambach, 2000). This should not come as a surprise, however, as the Precambrian occupies nearly 90% of Earth's history, a vast time interval for evolutionary development. As the pace of evolution can be rapid, the hindrance in the development of complex multi-

cellularity until over 3000 million years after the first hints of life presents numerous queries in our understanding of the history and nature of evolution on the Precambrian Earth. It is likely that a series of switches needed to be turned-on prior to the so-called "explosion" of metazoan life. Some of the previously proposed switches include, but are not limited to, internal genetic switches, such as the *Hox* gene complex, behavioral or ecological switches, including the evolution of macrophagy and increasing predation pressures, as well as numerous external switches, such as the increase of atmospheric oxygenation and changes in seawater chemistry (e.g., see Cloud, 1976; Peterson et al., 2005). One switch – or perhaps more accurately a compilation of interconnected switches – that was undoubtedly a prerequisite to metazoan evolution, is the reduction of seawater salinity and temperature along with the concurrent change in the potential for increased hydrospheric oxygen solubility (Knauth, 2005).

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If evolutionary change were dictated only by internal switches, one may suggest – with the presence of metaHox genes in primitive organisms such as sponges (Coutinho et al., 2003) – that complex multicellularity may have had the potential to arise much more swiftly than what we have observed in the fossil history of life. It seems applicable, therefore, to invoke external pressures, such as environmental restrictions, as levels of containment for life's impending evolutionary developments. Indeed, many traditional explanations for major Precambrian evolutionary events invoke specific thresholds in atmospheric oxygen levels; for instance, for both metabolic function as well as the biosynthesis of membrane-supporting sterols, the origin of Eukarya have been suggested to require at least 1% of the present atmospheric level (PAL) of oxygen (e.g., Jahnke and Klein, 1983; Han and Runnegar, 1992). Further, one of the more accepted explanations for the appearance of complex multicellularity in the Ediacaran is the increase in atmospheric oxygen content to a required threshold level (suggested to be between 1 and 10% PAL, depending on potential adaptation of Ediacara organisms to dysaerobic conditions) for metazoan respiration (e.g., Runnegar, 1982, 1991; Catling et al., 2005). While atmospheric oxygenation is undoubtedly important, as Knauth (2005) stated, “the level of oxygen dissolved in water is the real issue.” Seawater oxygen solubility is highly constrained by both ocean temperature and salinity, which likely fluctuated significantly over the course of the geologically- and environmentally-evolving Precambrian Earth.

The most continuous records of seawater temperature come from the complementary stories told by $\delta^{18}\text{O}$ and $\delta^{30}\text{Si}$ data, which have been collected and compiled from marine cherts (Knauth, 2005; Robert and Chaussidon, 2006). These isotopic data (along with other published data, for instance, see Perry, 1967; Knauth and Epstein, 1976; Knauth and Lowe, 1978) have been interpreted as evidence for a long-term cooling trend of the oceans through the Precambrian. From Knauth's (2005) estimates using a hypothetical 10–20 °C seawater temperature for Cenozoic oceans (see also Knauth and Lowe, 2003) and compiled oxygen isotope data, the Archean ocean surface temperature ranged from approximately 55–85 °C, which agrees with paleotemperature estimates of ~70 °C ascertained from silicon cycle models constructed from silicon isotopic data (Robert and Chaussidon, 2006). Furthermore, isotopic evidence of a hot Archean ocean has been linked to microbial evolution by studies of thermostability in early microbial proteins (Gaucher et al., 2008). However, it is important to note that recent isotopic data (Hren et al., 2009; Blake et al., 2010) challenge the commonly held view of a hot Archean world, pointing instead to a temperate and perhaps biologically active primitive ocean. Using a combined approach of oxygen ($\delta^{18}\text{O}$) and hydrogen (δD) isotopes from cherts of the ~3420 Ma Buck Reef Chert in South Africa, Hren et al. (2009) provide evidence suggesting that these cherts formed in equilibrium with waters below ~40 °C. Supporting these conclusions and adding yet another facet, Blake et al. (2010) ascertained $\delta^{18}\text{O}_\text{p}$ from phosphates of the ~3500–3200 Ma Barberton Greenstone Belt, South Africa. Due to their similarity with modern phosphates, these data again suggested a marine temperature of less than ~40 °C in the Archean and a developed biological phosphorus cycle (Blake et al., 2010). Thus, while seawater temperatures in the Archean have become debated in light of recent evidence, multiple estimates for Archean ocean salinity place the likely range between 1.2 and 2× that of modern ocean surface waters (Holland, 1978; Knauth, 1998). This elevated salinity may have persisted through the Archean due to the absence of continental cratons, as salt was not sequestered in halite deposits forming in evaporitic environments (Knauth, 2005). Furthermore, oxygen solubility in these early oceans was limited by elevated water temperatures and salinity, presumably resulting in a predominantly anoxic ocean full of anaerobic, thermophilic, and at least halotolerant if not

halophilic prokaryotes. One can easily imagine, then, that the physical and chemical properties, certainly including oxygen solubility, of this early ocean are far different than the ocean we know today.

In the Proterozoic Earth, on the other hand, seawater temperatures declined as continents became larger, more stable, and longer-lived. From Knauth's (2005) oxygen isotope dataset and Robert and Chaussidon's (2006) silicon isotopes, it is suggested that seawater temperatures reached typical Phanerozoic levels, an estimated 20 °C, by approximately 1200–800 Ma, during the assembly and near the break-up of the supercontinent Rodinia. During the break-up of Rodinia, the ocean temperatures of the Neoproterozoic Earth likely fluctuated dramatically, with global glaciation snowball events and warming periods during interglaciation; unfortunately, there seems to be a comparative lack of cherts during the Neoproterozoic and lower Paleozoic, which results in a scarcity of data for paleotemperature assessments during this biologically significant time interval (Knauth, 2005). Thus, seawater temperatures during the key geologic interval of metazoan origination and evolution is yet unresolved. Further, once in the Neoproterozoic, salt began precipitating out of seawater in large evaporite basins – such as the one described here. The resulting lowered oceanic salinity, in conjunction with the cooling of ocean temperatures associated with global glaciations, would allow dissolved oxygen to more readily accumulate in the oceans. This compilation of interconnected events not only limited the ecological ranges of anaerobic, thermophilic, halotolerant/-philic prokaryotes that had once dominated the oceans, but also may have resulted in enough dissolved oxygen in the ocean system to support metazoan respiration – and most importantly corresponded in time with the rise of complex metazoans, as evidenced by the earliest fossil representatives of metazoan life at ~632 Ma (Yin et al., 2007).

We provide additional data herein for seawater paleotemperature reconstruction of the terminal Neoproterozoic not based on isotopic analysis, but rather from halite fluid inclusion homogenization temperature assessments.

2. Halite fluid inclusions and geological setting of collected samples

Paleotemperature reconstructions using $^{18}\text{O}/^{16}\text{O}$ and $^{30}\text{Si}/^{28}\text{Si}$ methods vary largely with diagenetic influences and provide only relative estimates; whereas homogenization of primary halite fluid inclusions, by utilizing the cooling nucleation method, provides a direct means to evaluate seawater paleotemperature (Roberts and Spencer, 1995; Benison and Goldstein, 1999; Lowenstein et al., 1998; Satterfield et al., 2005; Liu et al., 2007). The cooling nucleation method of measuring homogenization temperatures in original, all-liquid, primary fluid inclusions has been successfully used in Permian, Pleistocene, and modern halites (Benison and Goldstein, 1999; Lowenstein et al., 1998; Satterfield et al., 2005). Maximum homogenization temperatures of fluid inclusions (T_{hmax}) in halite (from both laboratory-produced crystals and modern samples from Death Valley, CA) match the assessed brine temperatures during halite precipitation, and additionally correlate well with the average maximum air temperatures (Lowenstein et al., 1998). Therefore, maximum homogenization temperatures, calculated from a series of measured homogenization temperatures (T_{h}) in halite cores, reflect the maximum brine temperature and provide a close approximation for local air temperatures.

Primary halite has two crystal forms: cumulate crystals and chevron crystals in sedimentary surrounding (Roberts and Spencer, 1995; Benison and Goldstein, 1999; Lowenstein et al., 1998). Cumulate halite forms at the air–brine interface and can record the surface water temperature, which is directly comparable to air temperature (Lowenstein et al., 1998). Chevron halite, on the

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