Ocean redox conditions between the snowballs – Geochemical constraints from Arena Formation, East Greenland

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A B S T R A C T

The emergence of animal ecosystems is largely believed to have occurred in increasingly oxygenated oceans after the termination of the Sturtian and Marinoan glaciations. This transition has led to several hypotheses for the mechanism driving ocean oxygenation and animal evolution. One hypothesis is that enhanced weathering increased oceanic nutrient levels, primary productivity and organic carbon burial, and ultimately oxygenated the atmosphere and oceans. Another hypothesis suggests that an animal-driven reorganization of the marine geochemical cycles might have oxygenated the oceans. Through molybdenum (Mo), carbon (C), sulfur (S) isotopes and iron (Fe) speciation results from the Arena Fm, East Greenland, this study constrains ocean redox conditions during the Cryogenian, after the Sturtian deglaciation and before the major radiation of animals. Carbon and sulfur isotope stratigraphy is used to correlate the Arena Fm with other formations worldwide between the Sturtian and Marinoan glaciations (~720–635 Ma). The lower part of the Arena Fm (~25 m) consists of black shales deposited under locally euxinic conditions as evidenced by high proportions of highly reactive iron (Fe₃ox/Fe₃ = 0.38) and pyrite (Fe₃py/Fe₃hr > 0.7). These black shales display small Mo enrichments (<3 ppm) and low Mo/TOC compared to overlying shales and Phanerozoic euxinic sediments. The maximum δ⁹⁸Mo value is observed in the basal Arena Fm (1.5‰). Many samples display lower δ⁹⁸Mo than typical oceanic input fluxes, which can be explained by Mo isotope fractionation from a marine Mo pool with δ⁹⁸Mo ~ 1.3‰, similar to that inferred from other Cryogenic euxinic basins. The combination of low [Mo] and δ⁹⁸Mo suggests that widespread anoxia prevailed in the oceans at this time. Our data are consistent with most other studies from this time suggesting that ocean oxygenation was not linked to Snowball Earth deglaciation, but was delayed until animals effectively entered the scene.

1. Introduction

The Cryogenian Period (~720–635 Ma) was host to two massive glaciation events; the Sturtian glaciation (717–660 Ma) and the Marinoan glaciation (~640–635.3 Ma) (Bold et al., 2016; Rooney et al., 2014). The Cryogenian also likely hosted the earliest metazoan ecosystems (Erwin et al., 2011). For example, the first possible sponge fossils are from 659 to 636 Ma (Maloof et al., 2010), and biomarker evidence also suggests that demosponges were present in the pre-Marinoan oceans (Love et al., 2009). These findings raise the question of whether climatic responses to the global glaciations could have produced a habitable environment for metazoans, for example in the form of pervasive ocean oxygenation (Scott et al., 2008; Sahoo et al., 2012; Planavsky et al., 2010; Kendall et al., 2015; Frei et al., 2009). Conversely, the evolving marine animal and algal ecosystems could also have caused ocean oxygenation (e.g., zooplankton; eukaryote-dominated phytoplankton; Brocks et al., 2017; Butterfield, 2009a,b; Canfield; 2014; Erwin, 2015; Erwin et al., 2011; Lenten et al., 2014; Logan et al., 1995) or even dramatic climatic responses resulting in the global glaciations (Tziperman et al., 2011).

In this study, we investigate marine redox conditions in the oceans during the Cryogenian warm interval as recorded in the rock record of
the Arena Fm, East Greenland. At this time, metazoan ecosystems were still not abundant, and the Earth system relaxed from a long period of extreme cooling. Numerous other studies have explored ocean redox conditions from marine shales at this time, including the Datangpo Fm, South China (Feng et al., 2010; Li et al., 2012; Zhang et al., 2015), MacDonaldryggen Member of Ellobreen Fm, Svalbard (Canfield et al., 2008; Kunzmann et al., 2015), Tapley Hill, Australia, Twitya Formation (Sterling et al., 2015), NW Canada, and the Arena Fm, East Greenland (Canfield et al., 2008). These data suggest that anoxic and ferruginous conditions prevailed on the continental shelves with anoxic and sulfidic (euxinic) conditions at other localities. A few studies have attempted to constrain the globally integrated ocean oxygenation state from uranium isotopes in carbonates of the Taishir Fm, Mongolia (Lau et al., 2017) and Mo isotopes in euxinic shales from the Black River Dolomite, Tasmania and the Datangpo Fm (Cheng et al., 2017; Kendall et al., 2015). The molybdenum isotope composition of the euxinic shales suggest pervasive euxinia, whereas uranium isotopes in carbonates suggest that a short period of extensive ocean oxygenation followed the Sturtian glaciation before contracting towards a more anoxic ocean state during the Taishir anomaly (Lau et al., 2017). More constraints on the global oxygenation state and a better stratigraphic context are necessary to elucidate these discrepancies. Here, we provide chemostratigraphic data (δ^{13}C, δ^{34}S) together with local redox proxy data (Fe speciation, Fe/Al, Mo enrichments) and global redox proxy data (δ^{86}Mo) from the Arena Fm, East Greenland.

The Mo isotope system is a useful proxy to decipher the globally integrated oxygenation state of the oceans. Mo is soluble in oxygenated seawater with a slow removal rate that gives its long residence time in the ocean (440 kyr) and a globally uniform Mo isotope distribution in the oxic part of the oceans (Nakagawa et al., 2012; Miller et al., 2011; Kendall et al., 2017). This was probably also true in the Cryogenian where the predicted Mo residence time in the ocean, 100–200 kyr, was shorter than today, but still much greater than the oceanic mixing time scale (∼1.5 kyr) (Dahl et al., 2011; Miller et al., 2011; Sarmiento and Gruber, 2006). Molybdenum is efficiently scavenged into euxinic sediments deposited under anoxic bottom waters with more than 11 µM hydrogen sulfide (H_{2}S), where molybdate is converted to reactive thiomolybdate anions. In hydrographically restricted situations where the resupply of aqueous Mo from open-ocean seawater is slow, quantitative drawdown of Mo can subsequently occur, allowing the δ^{86}Mo of open ocean seawater to be recorded in the sediments (Neubert et al., 2008; Noordmann et al., 2015; Kendall et al., 2017). We discovered euxinic sediments near the base of the Arena Fm and applied the Mo isotope proxy to assess the global oxygenation state in the Cryogenian interglacial.

2. Geological settings

The Tillite Group (Cryogenian-Ediacaran) in East Greenland consists of five units, from below: Ulvesø, Arena, Storeelv, Canyon, and Spiral Creek formations. The Arena Fm (formerly Inter Tillite Beds) forms the interglacial unit between the two diamictite-bearing units Ulvesø and Storeelv (Stouge et al., 2011). Traditionally, the Tillite group has been correlated to the Polarisbreen Group in Svalbard based on comparable stratigraphic data (MacDonaldryggen Member of Ellobreen Fm, Svalbard (Canfield et al., 2008; Kunzmann et al., 2015), Tapley Hill, Australia, Twitya Formation (Sterling et al., 2015), NW Canada, and the Arena Fm, East Greenland (Canfield et al., 2008). These data suggest that anoxic and ferruginous conditions prevailed on the continental shelves with anoxic and sulfidic (euxinic) conditions at other localities. A few studies have attempted to constrain the globally integrated ocean oxygenation state from uranium isotopes in carbonates of the Taishir Fm, Mongolia (Lau et al., 2017) and Mo isotopes in euxinic shales from the Black River Dolomite, Tasmania and the Datangpo Fm (Cheng et al., 2017; Kendall et al., 2015). The molybdenum isotope composition of the euxinic shales suggest pervasive euxinia, whereas uranium isotopes in carbonates suggest that a short period of extensive ocean oxygenation followed the Sturtian glaciation before contracting towards a more anoxic ocean state during the Taishir anomaly (Lau et al., 2017). More constraints on the global oxygenation state and a better stratigraphic context are necessary to elucidate these discrepancies. Here, we provide chemostratigraphic data (δ^{13}C, δ^{34}S) together with local redox proxy data (Fe speciation, Fe/Al, Mo enrichments) and global redox proxy data (δ^{86}Mo) from the Arena Fm, East Greenland.

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residues after 1 h leaching with 10% HCl at 105 °C (Kristiansen, 2007). Decarbonated powders (~50 μg pure organic carbon) was transferred into tin capsules and subsequently combusted in an Euro EA Elemental Analyzer. Generated CO₂ was transferred into the IsoPrime mass spectrometer and measured carbon isotope ratios were calibrated using the inhouse standard AK ($\delta^{13}$C_{AK} = $-25.3\%$).

After correction for linearity, reproducibility of $\delta^{13}$C determinations was better than 0.1% (1 S.D.), whereas the uncertainty for $\delta^{18}$O was better than 0.15% (1 S.D., n = 10). All isotope ratio measurements were measured relative to the NBS-19 certified reference standard, and

Fig. 1. (A) Map of Greenland and the location of the sampling site, Ella Ø, marked with a red star. (B) Geological map of the Kong Oscar Fjord region indicating the position of Ella Ø (Ella Island). Ella Ø is part of the basin with Neoproterozoic-Ordovician deposits. The Caledonian orogeny caused subsequent tectonic disruptions. (C) Geological map of the field location near Bastionsbugt displaying Bed group 18 (brown), Bed group 19 (red), Ulvesø Fm (yellow), Arena Fm (grey), Storeelem Fm (green), Canyon and Spiral Creek Fm (purple), and Cambrian-Ordovician deposits. (D) Photo of field location at Ella Ø. Spiral Creek Fm (SC) and Ulvesø Fm (UL) can be observed directly on the photo. Modified from Kristiansen (2007). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Table 1
Geochemical data from the Arena shale formation, East Greenland. Total organic carbon (TOC) and total sulfur (TS) analyses were measured on newly crushed sample powders and compared to data obtained in 2006 by Kristensen (2007).

| Sample ID | Mo (ppm) | FeT (wt%) | FeCARB (wt%) | FeMAG (wt%) | FeOX (wt%) | FePY (wt%) | FeHR (wt%) | Al (wt%) | 1 SD Al (wt%) | Fe/Al (wt%/wt%) | Ca (wt%) | 1 sd Ca (wt%) |
|-----------|----------|-----------|--------------|-------------|-----------|-----------|-----------|----------|-------------|----------------|----------|--------------|
| K3-222    | 3.2      | 5.7       |              |             |           |           |           |          |             |                |          |              |
| K3-223    | 1.3      | 3.5       | 2.39         | -0.09       | 0.02      | 0.9       | 3.2       | 1.5      | 0.2         | 2.3            |          |              |
| K3-224    | 3.5      | 5.5       | 0.5          | 0.2         | 0.1       | 3.8       | 4.5       | 8.0      | 0.2         | 0.7            |          |              |
| K3-225    | 4.0      | 2.9       | -0.1         | 0.1         | 0.9       | 3.9       | 1.9       | 0.1      | 2.1         | 18             | 1.8      |              |
| Sample_ID | Mo (ppm) | FeT (wt%) | FeCARB (wt%) | FeMAG (wt%) | FeOX (wt%) | FePY (wt%) | FeHR (wt%) | Al (wt%) | 1 SD Al (wt%) | Fe/Al (wt%/wt%) | Ca (wt%) | 1 sd Ca (wt%) |
|-----------|----------|-----------|--------------|-------------|-----------|-----------|-----------|---------|---------------|----------------|----------|--------------|
| K3_226    | 2.4      | 4.6       | 1.4          | 0.0         | 0.0       | 2.2       | 3.6       | 2.9     | 0.1           | 1.6            | 12       | 1.9          |
| K3_228    | 2.9      | 5.6       | 0.4          | 0.1         | 0.0       | 3.8       | 4.4       | 8.0     | 0.2           | 0.7            | 1        | 1.8          |
| K3_229    | 1.5      | 2.3       | 1.4          | 0.1         | 0.0       | 1.3       | 2.6       | 2.4     | 0.1           | 1.0            | 21       | 1.9          |
| K3_230    | 5.6      | 0.4       | 0.1          | 0.0         | 0.0       | 4.0       | 4.6       | 8.3     | 0.2           | 0.7            | 1        | 1.8          |
| K3_231    | 5.0      | 0.7       | 0.1          | 0.0         | 0.0       | 2.2       | 3.0       | 4.7     | 0.2           | 1.1            | 9        | 1.9          |
| K3_232    | 0.9      | 6.0       | 0.4          | 0.2         | 0.1       | 5.4       | 6.1       | 8.9     | 0.2           | 0.7            | 1        | 1.8          |
| K3_233    | 0.9      | 6.3       | 0.4          | 0.3         | 0.1       | 3.1       | 3.9       | 8.1     | 0.2           | 0.8            | 2        | 1.8          |
| K3_234    | 1.4      | 6.1       | 0.3          | 0.3         | 0.0       | 3.2       | 3.9       | 8.2     | 0.2           | 0.7            | 2        | 1.8          |
| K3_235    | 1.4      | 6.4       | 0.5          | 0.5         | 0.0       | 2.1       | 3.1       | 7.7     | 0.2           | 0.8            | 2        | 1.8          |
| K3_236    | 0.2      | 6.6       | 0.5          | 0.7         | 0.1       | 0.4       | 1.6       | 6.6     | 0.2           | 1.0            | 5        | 1.8          |
| K3_237    | 0.2      | 5.5       | 0.3          | 0.6         | 0.2       | 0.3       | 1.3       | 8.6     | 0.2           | 0.6            | 2        | 1.8          |
| K3_238    | 0.2      | 9.4       | 0.5          | 0.6         | 0.1       | 0.2       | 1.4       | 7.7     | 0.2           | 1.2            | 2        | 1.8          |
| K3_239    | 0.6      | 6.9       | 0.4          | 0.5         | 0.1       | 2.8       | 3.8       | 7.8     | 0.2           | 0.9            | 2        | 1.8          |
| K3_240    | 0.6      | 6.3       | 0.4          | 0.5         | 0.1       | 1.9       | 2.8       | 8.3     | 0.2           | 0.8            | 1        | 1.8          |
| K3_241    | 0.6      | 6.1       | 0.3          | 0.4         | 0.0       | 0.1       | 0.7       | 4.6     | 0.2           | 1.3            | 8        | 1.9          |
| K3_242    | 0.6      | 6.6       | 0.4          | 0.7         | 0.1       | 0.0       | 1.2       | 9.0     | 0.2           | 0.7            | 1        | 1.8          |
| K3_243    | 0.6      | 6.6       | 0.4          | 0.7         | 0.1       | 0.0       | 1.2       | 9.0     | 0.2           | 0.7            | 1        | 1.8          |
| K3_244    | 0.6      | 6.5       | 0.2          | 0.9         | 0.4       | 0.7       | 2.2       | 10.8    | 0.3           | 0.6            | 1.2       | 1.8          |
| K3_245    | 0.6      | 5.8       | 0.3          | 0.8         | 0.7       | 0.7       | 2.5       | 9.0     | 0.2           | 0.6            | 1.0       | 1.8          |
| K3_246    | 0.6      | 9.1       | 0.6          | 0.7         | 0.1       | 0.2       | 1.7       | 7.2     | 0.2           | 1.3            | 1.8       | 1.8          |
| K3_247    | 0.6      | 6.9       | 0.3          | 1.1         | 1.1       | 0.5       | 3.0       | 8.1     | 0.2           | 0.8            | 1.0       | 1.8          |
| K3_248    | 0.6      | 7.4       | 0.6          | 1.1         | 0.4       | 0.4       | 2.5       | 8.3     | 0.2           | 0.9            | 1.6       | 1.8          |
| K3_249    | 0.6      | 7.9       | 0.4          | 0.9         | 0.7       | 0.4       | 2.3       | 8.6     | 0.2           | 0.9            | 1.8       | 1.8          |
| K3_250    | 0.6      | 5.7       | 0.2          | 0.4         | 0.3       | 0.5       | 1.5       | 8.1     | 0.2           | 0.7            | 2.2       | 1.8          |
| K3_251    | 0.6      | 6.3       | 0.4          | 0.9         | 0.6       | 0.4       | 2.3       | 6.9     | 0.2           | 0.9            | 2.5       | 1.8          |
| K3_252    | 0.6      | 5.8       | 0.3          | 0.7         | 0.3       | 0.2       | 1.5       | 6.7     | 0.2           | 0.9            | 3.1       | 1.8          |
| K3_253    | 0.6      | 7.8       | 0.3          | 0.3         | 0.0       | 1.4       | 2.0       | 6.0     | 0.2           | 0.7            | 3.7       | 1.8          |
| K3_254    | 0.6      | 4.2       | 0.3          | 0.3         | 0.0       | 1.4       | 2.0       | 6.0     | 0.2           | 0.7            | 3.7       | 1.8          |
| K3_255    | 0.6      | 6.3       | 0.3          | 0.5         | 0.1       | 0.6       | 1.5       | 7.3     | 0.2           | 0.9            | 2.8       | 1.8          |
| K3_256    | 0.6      | 6.7       | 0.3          | 0.5         | 0.1       | 1.2       | 2.1       | 8.1     | 0.2           | 0.8            | 1.6       | 1.8          |
| K3_257    | 0.6      | 6.7       | 0.3          | 0.5         | 0.1       | 1.2       | 2.1       | 8.1     | 0.2           | 0.8            | 1.6       | 1.8          |
| K3_258    | 0.6      | 6.3       | 0.3          | 0.4         | 0.0       | 0.7       | 1.5       | 7.8     | 0.2           | 0.8            | 2.3       | 1.8          |
| K3_259    | 0.6      | 6.5       | 0.4          | 0.5         | 0.1       | 0.8       | 1.8       | 8.0     | 0.2           | 0.8            | 1.7       | 1.8          |
| K3_260    | 0.6      | 7.5       | 0.4          | 0.5         | 0.1       | 1.6       | 2.7       | 8.0     | 0.2           | 0.9            | 2.2       | 1.8          |

* Measured using XRF. Calibrated using Ahms et al. (2017). STD = 0.10 wt%. 
3.2. Sequential Fe extraction

Iron speciation analyses were performed using a calibrated sequential extraction procedure for fine-grained continental sediments (Poulton and Canfield, 2005). Total Fe concentrations were obtained from rock powders by first roasting sediments 8 h at 520 °C, with subsequent digestion in 6 N HCl for 48 h (Aller et al., 1986). The concentrations of each iron pool were determined using a Perkin Elmer Atomic Absorption Spectrometer AAS-100 at the University of Southern Denmark. The analytical error for each extraction was less than 5%. Pyrite sulfur was extracted by chromium digestion (Canfield et al., 1986; Zhabina and Volkov, 1978), trapped as Ag₂S₂, and its concentration determined gravimetrically with a reproducibility of ± 5%.

3.3. Sulfur stable isotopes

Sulfur isotopic compositions were determined with a Thermo Analytical elemental analyzer, Flash EA 1112 series coupled via a ConFlo IV interface to a Thermo Delta V Plus mass spectrometer the University of Southern Denmark. The ³⁴S/³²S ratios are reported as δ³⁴S in terms of per mil relative to the Vienna Canyon Diablo Troilite (CDT) NIST 3134 is defined as 0.25‰ (Nägler et al., 2013). On this scale, modern seawater has δ⁹⁸Mo = 2.34 ± 0.10‰ (Goldberg et al., 2013). The long-term external reproducibility for δ⁹⁸Mo is ± 0.08‰, as estimated from repeat digestions of the USGS SDO-1 shale standard (2 S.D., n = 31, average δ⁹⁸Mo = 0.79%). Mo concentrations were calculated by isotope dilution using the ¹⁰⁶Mo/⁹⁵Mo ratio.

3.5. Elemental data

Elemental data for Al, Ca, and Fe were collected using hand-held X-ray Fluorescence (XRF) instrument Olympic DP-6000 at the University of Copenhagen. Loose ground powders were covered with 4.0 µm Prolene™ thin-film from Chemplex during analysis. Measurement time was 120 s for both heavy and light elements respectively. Samples were run with the reference standard PACS-3 before, during, and after analyses. XRF results were calibrated using method of Ahm et al. (2017). Estimated uncertainties (1SD) for Al, Ca, and Fe are 0.1–0.2 wt%, 1.8–1.9 wt%, and 0.1 wt%, respectively. Table 1 gives uncertainty estimates for each sample.

4. Results

4.1. Carbon and oxygen isotope composition of carbon

The lower part of the Arena Fm (375–390 m) shows a decreasing trend in δ¹³C³⁴CARB compositions from 1.4‰ to −4.5‰, reaching a constant minimum values of −5.1 ± 1.5‰ (average ± 1 SD, n = 9) for the rest of the section (Table 1, Fig. 2). Similarly, δ¹⁸OC⁴⁴CARB displays a decreasing trend (−3.8‰ to −14.1‰) in the lowermost section and constantly low values (−13.4 ± 0.6‰, 1 SD, n = 9) in the upper part of the section (Table 1, Fig. 2). In contrast, δ¹⁸OC⁴⁴COR displays a range of −33.7‰ to −30.4‰ in the lowermost interval (Table 1, Fig. 2).

4.2. Sulfur isotope composition of pyrite (δ³⁴S⁴⁴PY

The sulfur isotope compositions of pyrites in the lower Arena Fm display a narrow range of high δ³⁴S⁴⁴PY values (mean: 34 ± 3‰, 1 sd, n = 10), whereas the upper part of the section shows larger δ³⁴S⁴⁴PY variability from 10.8‰ to 64.8‰ (mean: 46 ± 12‰, 1 sd, n = 25) (Table 1 and Fig. 3). All samples display values above that of the typical
4. Sequential iron extraction (Fe\textsubscript{HR}/Fe\textsubscript{T})

The lowermost Arena Fm displays high Fe\textsubscript{HR}/Fe\textsubscript{T} > 0.38 characteristic of deposition in anoxic waters. The Fe\textsubscript{HR}/Fe\textsubscript{T} values systematically decline from 1.0 to ~0.3 in concert with Fe\textsubscript{PY}/Fe\textsubscript{HR} ratios from 0.9 to 0.1 (Table 1, Fig. 3). The decline parallels that of the TOC profile. This indicates that the lower part of the Arena Fm was deposited under an anoxic and sulfidic water column (euxinic conditions). In five of the twelve studied samples, the Fe-carbonate content dominates the highly reactive Fe pool. These samples also stand out with extremely high carbonate content (CaCO\textsubscript{3} > 20 wt%) (Table 1). The upper part of the Arena Fm (400–550 m) displays lower Fe\textsubscript{HR}/Fe\textsubscript{T} ratios oscillating between ~0.2 and ~0.5 with no discernable trends. On average, the highly reactive iron phases comprise of pyrite (38%), magnetite (32%), Fe-carbonates (19%) and a to a lesser extent Fe(III) oxides (12%) (Table 1). Importantly, Fe\textsubscript{PY}/Fe\textsubscript{HR} ratios remain below the euxinic threshold (0.7–0.8) although a few high values near the euxinic threshold occur at 554 m.

4.3. Total organic carbon (TOC) and total sulfur (TS)

Sedimentary TOC contents reach up to 3.0 wt% near the base of the Arena Fm (Fig. 3). Thereafter, TOC contents systematically decrease up-section (374–390 m) reaching consistently low TOC values averaging 0.16 ± 0.09 wt% at 400–550 m (1 SD, n = 26). The total sulfur content (TS) is higher in the lower part of the Arena Fm (mean: 2.5 ± 0.6 wt%) than in the upper part of the section (mean: 0.6 ± 0.7 wt%) (Table 1). The C/S (TOC/TS) weight ratios spans a wide range from 0.1 to 2.3 (mean 0.6 ± 0.6) with no systematic stratigraphic trends (Table 1). The average C/S ratios are well below that of modern normal marine shales (> 2.8) and modern euxinic sediments (> 2.8); and are comparable to Ordovician and Cambrian shales (0.5 ± 0.1) (Raiswell and Berner, 1985a,b). The relatively low C/S values have previously been ascribed to the absence of terrestrial refractory organic material and the absence of bioturbation that would have enhanced pyrite preservation (Raiswell and Berner, 1985a).

4.4. Fe/Al

The Fe/Al ratios in the samples range from 0.6 to 1.6 with no discernable trends (Table 1, Fig. 3). Importantly, the Fe/Al ratios remain above the 0.5 threshold indicative of sediment deposition in an anoxic marine basin (Lyons and Severmann, 2006). This result is robust despite the fact that samples taken from outcrop might have been affected by oxidative weathering. Fe/Al ratios are unaffected by the carbonate dilution in samples deposited closest to the underlying Ulvesø diamictite that have enhanced pyrite preservation (Lyons and Severmann, 2012, Poulton and Canfield, 2011). Deposition is indicated for anoxic samples with Fe/Al > 0.7 (Anderson and Raiswell, 2004; März et al., 2011). Proterozoic average Mo content is 21 ppm (Scott et al., 2008). The average δ\textsuperscript{98}Mo of modern day oceanic input is 0.7‰ (Archer and Vance, 2008). Error bars represent 2 SD (2 standard deviations). Ages of Sturtian and Marinoan glaciations from Bold et al., 2016; Macdonald et al., 2010; Rooney et al., 2014, 2015).

4.5. Sequential iron extraction (Fe\textsubscript{CARB}, Fe\textsubscript{OX}, Fe\textsubscript{MAG}, Fe\textsubscript{PY}, Fe\textsubscript{T})

Oceanic input (Phanerozoic values are < 8‰; Halevy et al. 2012), which is hard to reconcile with the fact that pyrite formation is associated with the preferential removal of lighter isotopes relative to overlying seawater.
average Mo/TOC ratio in the Arena Fm is 1.9 ± 1.0 ppm/wt% (1 SD, n = 11) with no clear stratigraphic trends. This is well below the average of Proterozoic shales (6.4 ppm/wt%) (Scott et al., 2008). However, one sample in the upper Arena Fm at 537 m is a single outlier that contains particularly low TOC content and 7.8 ppm Mo giving a Mo/TOC ratio of 24.5 ppm/wt%.

4.7. Molybdenum isotope composition (δ98Mo)

The δ98Mo profile shows a systematic decline upsection with values ranging from +1.47‰ to −0.26‰ in the lower part of Arena Fm (Table 1, Fig. 3). The δ98Mo values in the upper part are low and more variable at +0.12 to −0.86‰. These δ98Mo values are characteristically below that of average crustal rocks (~0.3‰) and average oceanic input (~−0.7‰), implying Mo isotope fractionation occurred during transport from the crustal source to the sediments (Fig. 4) (Kendall et al., 2017).

In a plot of δ98Mo vs 1/Mo (Fig. 4), it is clear that binary mixing between a Mo detrital source and an authigenic component does not explain all the chemical variability. However, δ98Mo and Mo/Al display a positive correlation in the lowermost part of the Arena Fm. From this, we infer that sediments of the lower Arena Fm contain an authigenic Mo component with high δ98Mo and modest Mo enrichments mixed with a very low δ98Mo and variably low Mo component. In addition, another authigenic component of Mo-depleted material with a negative δ98Mo value could contribute to the variability observed in the upper Arena shales. Assuming the detrital component in the lower Arena Fm has a composition similar to the average of bulk samples in the upper part of the Arena Fm; i.e., (Mo/Al)detr = 0.04 ± 0.02 (1 SD, n = 5) and δ98Mo,detr = −0.45 ± 0.42‰ (1 SD, n = 5), we estimate the Mo isotope composition of the “high Mo” authigenic component by subtracting the detrital component:

\[
δ_{98}^{Mo,\text{AUTH}} = [δ_{98}^{Mo} − (1-f_{\text{auth}})δ_{98}^{Mo,detr}]/f_{\text{auth}}
\]

where \(f_{\text{auth}} = 1 − (\text{Mo/Al})_{\text{detr}}/(\text{Mo/Al})\).

Our data shows that δ98Mo,AUTH systematically declines up-section with the highest values in the lower Arena Fm at 1.57 ± 0.06‰ to −0.53 ± 0.06‰. Uncertainty estimates for δ98Mo,AUTH represent propagated errors assuming the variables in Eq. (1) are independent.

5. Discussion

5.1. The age of the interglacial deposits in East Greenland

Chemo- and lithostratigraphic correlations provide an age of the Arena Fm suggesting the Ulvesø and Storeelv diamicties are of Sturtian and Marinoan age, respectively (Hoffman et al., 2012; Fairchild et al., 2016). Previous chemostratigraphic studies mainly focused on the Polarsibreen Group of NE Svalbard (Halverson et al., 2004; Hoffman et al., 2012; Fairchild et al., 2016) although Kristiansen (2007) provided δ13C data for the Tillite Group and Eleonore Super Group. A negative δ13C anomaly is recorded in the pre-glacial Bedgroup 19/20 (Kristiansen, 2007; Stouge et al., 2011). Early studies attributed this excursion to the
late Cryogenian Trezona anomaly implying that the Sturtian glaciation is not preserved in East Greenland (Halverson et al., 2004; Kristiansen, 2007; Stouge et al., 2011). In these models, the MacDonaldryggen Mb and Arena Fm formed as part of the same basin during the Marinoan glaciation by meltwater plumes underneath an ice-covered ocean (Halverson et al., 2004). However, the more recent recognition of the Islay negative anomaly in pre-Sturtian (Tonian) strata worldwide has led to a re-assessment of the negative δ13C anomaly in Bedgroup 19/20, supporting a Sturtian age of the Ulvesø Fm (Halverson et al., 2010; Hoffman et al., 2012). Given that the Storeeiv diamicite represents the Marinoan ice age as determined in Hoffman et al. (2012), the Arena Fm must have been deposited during the Cryogenian warm interval (Halverson et al., 2010; Hoffman et al., 2012; Fairchild et al., 2016). The δ13C and δ34S data provided in this study are consistent with this view, as explored below.

5.1.1. Characteristically extreme δ34S values in pyrite

The sulfur isotope composition of marine pyrites, δ34Spy, rarely exceeds average Phanerozoic oceanic input (< 8‰) (Haley et al., 2012) simply because pyrite is the major sink for reduced sulfur that preferentially removes the lighter sulfur isotopes from the marine sulfate reservoir (Canfield, 2004). Nevertheless, several Cryogenian interglacial deposits including MacDonaldryggen Fm, Datangpo Fm, Aralka Fm, Twitya Fm, and Tapley Hill Fm display extremely heavy values, at ∼30-60‰, similar to those observed in the Arena Fm (Feng et al., 2010; Li et al., 2012; Ries et al., 2009). Although extremely heavy pyrite δ34S values are not a worldwide signature in the ocean (e.g. Oman), it is clear that an extremely sulfate-depleted ocean is required to achieve these values (Ries et al., 2009). This property is typical, although not unique, for the Cryogenian warm interval oceans (e.g. Nama Group, 549–545 Ma) (Gorjan et al., 2000; Ries et al., 2009). Therefore, the δ34S data in the Arena Fm is consistent with previous chemo- and lithostratigraphic correlations.

5.1.2. δ13CORG profile of the Arena Fm

Several lines of evidence suggest that the δ13CORG profile in the Arena Fm shales is not representative of contemporaneous seawater due to the negligible carbonate content, its potentially authigenic origin and/or diagenetic overprint. The negative δ13CORG trend from +1.4 to −4.5‰ in the lower Arena Fm is not coupled to changes in δ13CORG as would be expected if both carbonates and marine biomass precipitated −8‰ (Kunzmann et al., 2015). In these models, the MacDonaldryggen Mb, Datangpo Fm, Aralka Fm, Twitya Fm, and Tapley Hill Fm display extremely heavy values, at ∼30-60‰, similar to those observed in the Arena Fm (Feng et al., 2010; Li et al., 2012; Ries et al., 2009). Although extremely heavy pyrite δ34S values are not a worldwide signature in the ocean (e.g. Oman), it is clear that an extremely sulfate-depleted ocean is required to achieve these values (Ries et al., 2009). This property is typical, although not unique, for the Cryogenian warm interval oceans (e.g. Nama Group, 549–545 Ma) (Gorjan et al., 2000; Ries et al., 2009). Therefore, the δ34S data in the Arena Fm is consistent with previous chemo- and lithostratigraphic correlations.

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5.2. Local redox conditions in the Arena basin

The local redox conditions in the Arena basin are similar to several other anoxic basins during the Cryogenian interglacial, namely MacDonaldryggen Mb, Datangpo Fm, Twitya Fm and Black River Dolomite (Feng et al., 2010; Li et al., 2012; Zhang et al., 2015; Canfield et al., 2008; Kunzmann et al., 2015). Anoxic and non-sulfidic (ferruginous) bottom water conditions have been inferred for the MacDonaldryggen Member and the Twitya Fm (Canfield et al., 2008; Kunzmann et al., 2015), whereas the Black River Dolomite are consistently euxinic (Kendall et al., 2015) and the lower part of the Datangpo Fm were deposited under euxinic and/or ferruginous conditions (Feng et al., 2010; Li et al., 2012; Cheng et al., 2017).

The Fe speciation and Fe/Al data constrain redox conditions in the bottom waters of the local marine depositional environment. Highly reactive Fe phases, FeIBR, comprise carbonates (FeCORG), ferric oxides (FeOx), magnetite (FeMAC), and pyrite (FePY) that characteristically exceed 38% of the total iron, FeIBR/FePy > 0.38, in marine settings with anoxic bottom waters. Lower FeIBR/FeT ratios may suggestoxic conditions, although high depositional rates of detrital material also produce deceptively low FeIBR/FePy (Poulton and Raiswell, 2002; Raiswell and Canfield, 1998). The FePy/FeIBR ratio distinguishes between an anoxic, non-sulfidic (potentially Fe2+ containing) water column and a sulfidic water column (Poulton and Canfield, 2001). Sediments deposited under anoxic and sulfidic (euxinic) waters contain high FePy/FeIBR ratios above ∼0.7 for Phanerozoic sediments (Anderson and Raiswell, 2004; März et al., 2011).

Fe speciation data in the lower Arena Fm (375–390 m) imply predominantly euxinic conditions (Fig. 6). However, euxinic samples are interbedded with five samples with FePy/FeIBR < 0.7 in which Fe-carbonate dominates the highly reactive Fe pool (Fig. 6). These five samples all display low Al (1.5–4.7 wt%), high Ca (9.7–21 wt%), high total carbon contents (4.3–10.5 wt%), and high FeCORG (0.7–2.9 wt%) (Fig. 6). These features suggest inorganic carbonate precipitation with prevalent Fe substitution. The temporal regularity of these interbedded samples suggests cyclical precipitation events. It is likely that such surges in precipitation of carbonates radically augmented sedimentation rates in those temporal intervals (Clarkson et al., 2014). Therefore, we infer that the lower Arena Fm (375–390 m) represents a primarily euxinic environment.

In the upper Arena Fm, FeIBR/FeT values are close to the 0.38 threshold (0.36–0.48) suggesting that either oxic or anoxic/ferruginous
bottom water conditions prevailed in the basin (Fig. 6). All samples of the Arena Fm display higher Fe/Al ratios (0.93 ± 0.36) than average shales and average river-suspended particulates (0.5–0.56). The upper Arena Fm is also substantially enriched (Fe/Al = 0.86 ± 0.19) indicative of ferrous iron transport within anoxic bottom waters of the Arena basin. However, the Arena shales display a lower Al/Ti mass ratio (14.9 ± 3.7) than average upper continental crust (24.8), and are more similar to mafic rocks of the Franklin Large Igneous Province (Al/Ti = 8–20) with higher Fe/Al mass ratio (0.8–1.5). This observation suggests that the detrital component in the Arena shales could contain higher Fe/Al than typical upper continental crust sediments (Cox et al., 2015; Cox et al. 2015; Dostal et al., 1985; Wedepohl, 1995). In combination with the muted Fe_{bio}/Fe_{t} = 0.30 ± 0.11 enrichments (this study) and limited Fe isotope fractionation in bulk rock samples from MacDonald-dryagen and Arena Fm, we suggest that the detrital Fe component overwhelmed any shelf-to-basin reactive iron shuttle (Kunzmann et al., 2017a). Therefore, the Fe-based proxies are not decisive for the upper Arena Fm and it could have been deposited either in oxic settings or under anoxic/ferruginous bottom waters with a mafic detrital component.

### 5.3. Basinal restriction

Today, the Mo scavenging process gives rise to Mo-TOC co-variation in euxinic sediments with slopes between ~5 and ~45 depending on water exchange between the basin and the open ocean. The highly restricted Black Sea, for example, yields 9-fold lower Mo/TOC ratios than the more open setting of the Saanich Inlet (Algeo and Lyons, 2006). The observed Mo/TOC slopes in the euxinic Arena shales, at 1.3 ± 0.3 ppm/wt%, are substantially lower than observed for any modern euxinic setting (Fig. 7), indicative of either an extremely restricted basin and/or lower Mo availability in the Cryogenian oceans. The Mo/TOC slope is similar to that of ~640 Ma organic mudrocks of the Black River Dolomite in Tasmania (mean: 3.5 ± 1.1 ppm/wt%, 1 SD, n = 10; Kendall et al., 2015) and 3 to 15-fold lower than in the ~663–654 Ma Datango sections, implying that the Arena Fm and the Black River Dolomite were deposited in more restricted basins (Cheng et al., 2017). However, the Greenland-Svalbard basin was laterally extensive, contained characteristic marine acritarchs, with strontium isotope compositions matching similar age marine carbonate deposits (Halverson et al., 2004; Hoffman et al., 2012; Kunzmann et al., 2015; Fairchild et al., 2016). Therefore, we infer that the Greenland-Svalbard basin, although restricted, still exchanged seawater with the open ocean.

Importantly, Mo enrichments are also low in euxinic sediments from the more open-marine euxinic Datango basin (~30 ppm) relative to modern open-shelf euxinic settings (~130 ppm) suggesting ~4-fold lower Mo concentration in the Cryogenian ocean than today (Cheng et al., 2017; Li et al., 2012). The markedly lower Mo concentrations in the Arena Fm (0.1–7.8 ppm), even compared to the Datango Fm, could indicate that hydrogen sulfide levels in the basal Arena Fm (375–390 m) were too low for efficient conversion of molybdate to reactive thiomolybdate. We note that some sulfidic sediments deposited under anoxic conditions do produce Mo-TOC co-variation with a shallow slope similar to the Arena Fm (Poulson Brucker et al., 2009). Therefore, we cannot rule out that low sedimentary Mo/TOC was produced in euxinic basins where bottom water H_{2}S concentrations never exceeded the 11 µM threshold.

### 5.4. Reconstructing global redox conditions

The marine Mo isotope budget varies as a function of the global ocean oxygenation state. The major source of oceanic Mo is in the form of MoO_{4}^{2−} via rivers with an average δ^{98}Mo value of 0.7‰ (Archer and Vance, 2008). In oxygenated waters, molybdenum isotope fractionation occurs when Mo is adsorbed onto Mn-oxides resulting in an isotopic difference of −3‰ between the Mn-oxide containing sediments and seawater. In euxinic settings where H_{2}S concentrations exceed 11 µM and Mo is converted to reactive thiomolybdate ions (MoO_{4}^{2−}, S_{x}^{2−}), molybdenum may be quantitatively removed with no net isotopic difference expressed between the sediments and seawater source (e.g., see Kendall et al., 2017 for a recent review). Another important Mo sink is settings of intermediate redox conditions, where Mo is eventually adsorbed onto Mn-oxides, such as ferrihydrite, hematite, magnetite, and goethite, in the oxic zone followed by reductive dissolution may contribute to a variable and smaller isotope offset from seawater; e.g., ~0.8‰ (Poulson Brucker et al., 2009; Goldberg et al., 2012). Isotope fractionation between sediments and seawater always cause preferential removal of isotopically lighter Mo isotopes leaving the modern ocean with a uniform and isotopically heavy δ^{98}Mo composition of 2.34 ± 0.10‰ (Kendall et al., 2017).

Euxinic sediments have the potential to record the δ^{98}Mo isotope composition of penecontemporaneous ocean seawater and constrain the globally integrated extent of Mo-sulfide removal into seafloor sediments (Arnold et al., 2004; Dahl et al., 2011). Effective transfer of seawater δ^{98}Mo values into the sedimentary rock record depends on enough hydrogen sulfide available for quantitative thiomolybdate formation in the local basin, and on hydrographic conditions that allow enough time for quantitative removal to occur (Dahl et al., 2010a; Neubert et al., 2008). Currently, no independent control measures can be taken to ensure that a sedimentary rock captures the ‘seawater’ value (Dahl and Wirth, 2017). Molybdenum isotope fractionation occurring in the local basin produces sediments with lower δ^{98}Mo than overlying seawater. Therefore, the highest δ^{98}Mo value in a section is typically used as a minimum estimate of seawater δ^{98}Mo (Dahl et al., 2010b), but we can also assess seawater δ^{98}Mo if we understand the magnitude of Mo isotope fractionation associated with sediment burial in the Arena Basin. We explore both of these approaches further below.

#### 5.4.1. Mo-isotope fractionation throughout Arena Fm

The highest δ^{98}Mo value in the Arena Fm is 1.5‰, occurring in euxinic sediments close to the Ulvøess (Sturtian-age) tillite. Therefore, this is a lower estimate of seawater δ^{98}Mo at this time in the post-Sturtian oceans. Several of the euxinic samples in the lower part of the...
Arena Fm carry authigenic Mo enrichments with much lower $^{\delta^{98}}$Mo values ($> -0.53\%$) than the oceanic Mo inventory can achieve as a whole ($\pm 0.7\%$) (Kendall et al., 2017). Hence, Mo isotope fractionation occurred most of the time when dissolved Mo was transported into the Arena basin and/or within it. Indeed, we also know that local Mo isotope fractionation must have taken place when bottom water H$_2$S levels dropped below the threshold of thiomolybdate formation during the transition from euxinic to ferruginous conditions in the Arena basin. At this time, $^{\delta^{98}}$Mo systematically decreased from 1.47% to $-0.26\%$. Thus, decreasing bottom water H$_2$S would imply that molybdate conversion to thiomolybdate was incomplete and that another Mo burial pathway took over or even dominated while conditions were still (perhaps mildly) euxinic.

The upper Arena shales display low and variable $^{\delta^{98}}$Mo isotope compositions (0.12% to $-0.86\%$) and very low sedimentary Mo contents that we ascribe to Mo burial associated with Fe-oxide precipitation in an anoxic and ferruginous environment. Molybdenum adsorbs onto magnetite, ferrhydrite, goethite, and hematite with distinct magnitudes of isotopic differences at 0.83 $\pm$ 0.60%, 1.11 $\pm$ 0.15%, 1.40 $\pm$ 0.48%, and 2.19 $\pm$ 0.54% respectively (Goldberg et al., 2009). We propose that Mo co-precipitated with ferrhydrite (Fe$_2$O$_3$·0.5H$_2$O) from the water column and subsequently redissolved in the sediment pore fluids, before some fraction was captured in magnetite, goethite, pyrite, and Fe-carbonate during early diagenetic Fe reduction. Hematite is not a product of early diagenesis and formed later. Molybdate ions (MoO$_4^{2-}$) from seawater were also delivered by diffusion into the pore fluids, where Mo fixation in insoluble Fe phases took place. This process would only dilute the isotopic offset between the sediments and the open ocean. Ultimately, the maximal isotopic difference between sediments and the seawater source also depends on how large a proportion of Mo is escaping back into the water column by diffusion (Clark and Johnson, 2008). The diagenetic Mo capture efficiency in the sediments depends on 1) the depth at which the Mo-shuttle (e.g. ferrhydrite) dissolves and liberates Mo back to interstitial pore-spaces, 2) the depth at which Mo re-precipitates with other Fe oxide phases (e.g. magnetite/goethite/pyrite/Fe-carbonate) and 3) the Mo affinities of the involved mineral phases, and 4) the availability of Mo surface sites on the final host phase. For example, we expect little diffusional Mo loss if ferrhydrite dissolution occurs deep inside the sediments, fresh Fe oxides immediately form at the same depth, and there are plenty of surface sites available for Mo capture in the sediments. In this situation, essentially all Mo will be captured in the magnetite/goethite/pyrite/Fe-carbonate inside the sediments and there will be no further isotope fractionation associated with the final Mo fixation step. Alternatively, if ferrhydrite dissolves near the sediment-water interface and Mo is liberated in the form of MoO$_4^{2-}$, and another reactive Fe phase instantaneously forms, then we expect approximately half of the Mo (e.g. the fraction that diffuses in a downwards direction) to be eventually incorporated into freshly formed insoluble Fe phases, whereas the rest is lost to the water column. In this case, the magnitude of isotope offset between sediment and dissolved MoO$_4^{2-}$ in the pore fluids will be $-\sim$ half of the isotope discrimination between reactant and products; that is $-0.41 \pm 0.30$ to $-0.70 \pm 0.24\%$ for magnetite and goethite, respectively. Adding the isotope fractionation associated with ferrhydrite precipitation, we calculate that sediments would carry variable $^{\delta^{98}}$Mo values that are maximally between 1.52 $\pm$ 0.45% and 1.81 $\pm$ 0.39% below that of contemporaneous seawater (calculations: 1.11 $\pm$ 0.15% + 0.41 $\pm$ 0.30% and 1.11 $\pm$ 0.15% + 0.70 $\pm$ 0.24%). Hence, both the absolute $^{\delta^{98}}$Mo values and the variability between samples in the ferruginous Arena shales, at $-0.45 \pm 0.42\%$, are fully explained by differences in the Mo removal pathway within the Arena basin and open ocean seawater at a constant steady state $^{\delta^{98}}$Mo value between 1.07 $\pm$ 0.61% and 1.36 $\pm$ 0.57% (calculations $-0.45 \pm 0.42\%$ + 1.52 $\pm$ 0.45%$/-0.45 \pm 0.42\% + 1.81 \pm 0.39\%)$. Clearly, this estimate is uncertain in part because we do not know the exact Mo burial pathway, or the efficiency of Mo sequestration during early diagenesis and because the magnitude of Mo isotope fractionation during pyrite and Fe-carbonate precipitation in pore fluids are also not known. Nevertheless, our model and calculations demonstrate that Cryogenian seawater may well have had $^{\delta^{98}}$Mo similar to earlier Neoproterozoic oceans (in contrast to the view held in Baldwin et al., 2013), and we conclude that seawater $^{\delta^{98}}$Mo was most likely well below that of the modern ocean.

5.4.2. Estimating the globally-integrated ocean oxygenation state, ~660–636 Ma

The lower Arena Fm contains a maximum $^{\delta^{98}}$Mo composition of $-1.5\%$, and we provide a model-dependent and more uncertain estimate that the upper part of the Arena Fm was deposited with seawater $^{\delta^{98}}$Mo = 1.1–1.3% ($\pm$ 0.6%). The contemporaneous lower part of the euxinic Datangpo Fm and the Black River Dolomite display maximal $^{\delta^{98}}$Mo values at $-1.1\%$ (Kendall et al., 2015; Cheng et al., 2017). Hence, the best estimates for the $^{\delta^{98}}$Mo value for the Cryogenian warm interval oceans falls in the 1.1–1.5% range. Low Mo concentrations (0.9–3.5 ppm) and Mo/TOC (1.5–3.3) in the euxinic part of the Arena Fm compared to modern environments is also a general trend observed in the Datangpo Fm and other pre-Marinoan samples (Scott et al., 2008; Dahl et al., 2010a,b; Cheng et al., 2017). This finding implies that low Mo availability in the global ocean environment is a characteristic feature of the Cryogenian. In combination with the estimated ~4-fold lower Mo concentration in Cryogenian seawater compared to modern oceans, our observations indicate a global ocean with extensive euxinia occurring after the Sturtian glaciation.

According to a model for the Mo cycle in the ancient oceans (Dahl et al., 2011), there is no need for an oxygenated deep ocean (today Mn-oxides in the deep ocean accounts for 30–50% of global Mo burial). Model predictions show that euxinic conditions covering 0.1–1.0% of the seafloor would lead to oceanic $^{\delta^{98}}$Mo of 1.1–1.5% and 3–6-fold lower Mo concentrations than the modern ocean (Dahl et al., 2011). This is a 2–20 fold greater extent of euxinic seafloor area compared to modern oceans and similar to the pre-Sturtian ocean state (Dahl et al., 2011). By inference, sulfidic waters masses extended beyond restricted fjords and occurred in continental shelves during the Cryogenian warm interval, which is also exemplified in the Arena, Datangpo and Black River Dolomites basins. Still, anoxic and ferruginous waters were likely even more pervasive on the continental shelves and the deep ocean could well have been anoxic (Canfield et al., 2008; Dahl et al., 2010b; Sperling et al., 2015).

Molybdenum isotopic evidence from euxinic shales in Greenland, China and Australia suggests that euxinic ocean conditions still prevailed after the Sturtian glaciation. Hence, we find no evidence that the thawing of the Sturtian Snowball led to massive ocean oxygenation (Feng et al., 2010; Li et al., 2012; Zhang et al., 2015; Cheng et al., 2017; Kendall et al., 2015). This observation differs from findings in recent studies of U isotopes in the Taishir Fm, Mongolia. These studies show a contrasting development in the global ocean oxygenation state after the Sturtian glaciation (Lau et al., 2017). Mass balance analyses of the $^{\delta^{238}}$U results showed a period of ocean oxygenation immediately after the Sturtian glaciation followed by a return to anoxia at the onset of the negative Taishir $^{\delta^{13}}$CARB excursion (Lau et al., 2017). The marine U cycle is sensitive to sediment burial in anoxic settings, whereas the Mo cycle depends on hydrogen sulfide present in the anoxic waters (Dahl et al., 2014). Therefore, it is hard to reconcile how the marine Mo cycle could be in a more sulfidic anoxic state when the marine U cycle was in an oxygenated state.

We offer two solutions to the problem. First, it is theoretically possible that the Mo and U data do not overlap in time. The positive $^{\delta^{13}}$CORG transition in the Arena Fm may indicate that it was deposited only during the anoxic period of the Tayshir excursion, missing the oxygenated period leading up to it. This would require that the Arena Fm, Datangpo and Black River Dolomite only represented the later
pervasively anoxic period of the interglacial interval. Higher chrono-
and chemostратigraphic resolution of the respective sampling locations can clarify whether these shale formations contain the earlier part of the Cryogenian interglacial. Secondly, the lower Taishir Fm may carry a false oxygenation signature prior to the Taishir $^{3}^{13}$C anomaly; for example by incorporation of reduced U (as indicated by elevated U contents in this part of the two studied sections) (Lau et al., 2016). This behavior is exemplified in some Late Cambrian argillaceous limestones (Dahl et al., 2014). To test this idea, we recommend further uranium isotope studies from coeval limestone sections elsewhere in the world. These arguments would imply that anoxic ocean conditions were pervasive throughout the Cryogenian warm interval.

5.5. Ocean redox state and animal evolution

According to observations in this study, ocean oxygenation did not follow the termination of the Sturtian glaciation. However, chemos-
stratigraphic constraints are needed to correlate the Arena Fm with the Ts--shir Fm and additional U isotope studies are needed to validate the global oxygenation episode recorded in the Taishir Fm (Lau et al., 2016). After the Marinoan glaciation, several studies suggest episodes of widespread ocean oxygenation (Kendall et al., 2015; Sahoo et al., 2012) superimposed on a predominantly anoxic Ediacaran ocean state (Canfield et al., 2008; Sahoo et al., 2016; Sperling et al., 2015). Therefore, we ask what made the post-Marinoan interval different from the post-Sturtian deglaciation?

Marine oxygen levels are determined by the balance between $O_2$ supply from the atmosphere and $O_2$ consumption within the oceans. Atmospheric $O_2$ levels are linked to continental weathering via oceanic nutrient supply, marine primary productivity and organic carbon burial (Lenton et al., 2014). Model predictions suggest global weathering rates increases ~10 fold in the aftermath of a Snowball glaciation (Le Hir et al., 2009). This increase could fertilize the oceans and thereby increase marine $O_2$ consumption, and at the same time sustain more or-
geanic carbon burial that would increase atmospheric $pO_2$ if sustained over million year time scales (Lenton et al., 2014; Mills et al., 2011; Planavsky et al., 2010). Our data indicate that marine oxygen consumption kept pace with any putative rise in dissolved $O_2$ levels in the surface ocean resulting from atmospheric oxygenation after the Sturtian glaciation. Nevertheless, we cannot exclude that ocean oxygenation occurred as a delayed response to increased weathering rates during a subsequent period of time not represented by the Arena Fm.

In contrast, biological driving factors could well reorganize the marine carbon cycle and cause widespread ocean oxygenation. During the Late Neoproterozoic, eukaryotic algae replaced cyanobacteria as the most abundant primary producer and animals with bilaterian, diplo-
lastic, and poriferan body plans appeared (Erwin et al., 2011; Knoll, 2014). The larger organisms might have driven an internal re-
origination of the marine carbon cycle necessary to drive ocean oxy-
genation in at least two ways. For example,

1) Filter-feeding organisms (e.g. sponges, cnidaria) emerging in the Tonian and Cryogenian oceans were capable of clearing the water column of organic material, especially the smallest picoplankton and dissolved organic matter (Peterson et al., 2004; Love et al., 2009; Maloof et al., 2010; Erwin et al., 2011; Sperling et al., 2015 (b)). Thus, sponges might have removed dissolved organic matter from the water column and thereby reduced the oxygen demand in the shallow oceans, and also allowing light to penetrate deeper into the water column giving a selective advantage for larger eukaryotic algae (Butterfield, 2009a,b; Lyons et al., 2014; Lenton et al., 2014).

In turn, this process could have removed a large $O_2$ sink within the oceans and thereby oxygenated the oceans (Butterfield, 2009a,b; Lenton et al., 2014). It might even have generated a positive feed-
back loop if filter-feeding organisms invaded new areas of the sea-
floor or became more efficient in increasingly oxygenated oceans

(Lenton et al., 2014).

2) The size of zooplankton and their fecal pellets is another a key factor dictating organic carbon export to the sediment-water interface (Logan et al., 1995). An increase in fecal pellet size would cause an increase in the sinking velocity of organic particles through the water column. Due to greater decomposition depths, the ocean surface became increasingly oxygenated and hydrogen sulfide pro-
duction was confined to the ocean floor rather than in the water column (Canfield, 2014; Lenton et al., 2014; Logan et al., 1995).

In both cases, atmospheric oxygenation preceded ocean oxygena-
tion. The Sturtian and Marinoan glaciations did not trigger animal evolution via atmospheric oxygenation. Instead, marine oxygen levels must have already been sufficiently high to meet early animal respira-
tory needs (Mills and Canfield, 2014). Rather, the Cryogenian icehouse could have promoted animal evolution in other ways (e.g., Boyle et al., 2007), so that the delayed biological reorganizations by animals promoted further ocean oxygenation and reinforced the es-
establishment of metazoan ecosystems in the Ediacaran and Cambrian.

6. Conclusion

This study investigated the chem stratigraphic and ocean redox conditions of the Arena Fm, East Greenland. Based on a correlation between $^{3}^{13}$C chem stratigraphy and extremely heavy $^{3}^{13}$S, the depo-
sites of the Arena Fm represent the warm interval between the Sturtian and Marinoan glaciations. The $^{3}^{13}C_{CARB}$ record has been di-
ag enically altered by $^{3}^{13}$C-depleted meteoric water and dolomitizing fluids. Iron-based proxies show that the depositional environment was anoxic, and transitioned from euxinic conditions near the base of the succession to furrowous conditions up-section. Very low Mo con-
centrations and a Mo–TOC correlation with a shallow slope in the eux-
inic interval suggest that the Arena basin was hydrographically re-
stricted and that dissolved Mo was ~4-fold lower in Cryogenian seawater compared to the present day. The maximum $^{3}^{13}$Mo value recorded in the euxinic Arena shales is 1.5‰, which is well below that of the modern oceans (2.3‰). A decreasing trend of $^{3}^{13}$Mo upsection is correlated with an inferred decline in the concentration of bottom water $H_2S$ in the basin, culminating in strongly fractionated $^{3}^{13}$Mo values (~0.86‰). These very low values are likely due to the removal of Mo from the water column in association with Fe-oxide minerals, and/or incomplete thiomolybdate conversion at low $H_2S$ availability (< 11 μM). In combination with the highest $^{3}^{13}$Mo values observed in contemporary euxinic basins (Datangpo Fm, China, and Black River Dolomites, Tasmania), our results indicate that the deep oceans were not fully oxygenated and that a ~0.1-1% fraction of the global seafloor remained pervasively euxinic in the aftermath of the Sturtian glaciation. Therefore, we infer that if enhanced weathering occurred, it did not significantly perturb the extent of euxinia in the oceans. These observations contrast with global redox studies in carbonates of the Taishir Fm, Mongolia. Further global ocean redox constraints and better chem stratigraphic resolution are necessary to settle discrepancies in interpretation of the redox conditions in the Cryogenian warm interval and the link between Snowball glaciations and ocean oxygenation.

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