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Variations in glacial and interglacial marine conditions over the last two glacial cycles off northern Greenland

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A B S T R A C T

Five sediment cores from the Lomonosov Ridge and the Morris Jesup Rise north of Greenland show the history of sea-ice coverage and primary productivity over the last two glacial cycles. Variations in Manganese content, benthic and planktonic foraminifera, bioturbation, and trace fossil diversity are interpreted to reflect differences in sea-ice cover and sediment depositional conditions between the identified interglacials. Marine Isotope Stage (MIS) 1 and MIS 2 are represented by thin (<<5 cm) sediment units while the preceding interglacial MIS 5 and glacial MIS 6 are characterized by thick (10−20 cm) deposits. Foraminiferal abundances and bioturbation suggest that MIS 1 was generally characterized by severe sea-ice conditions north of Greenland while MIS 5 appears to have been considerably warmer with more open water, higher primary productivity, and higher sedimentation rates. Strengthened flow of Atlantic water along the northern continental shelf of Greenland rather than development of local polynyas is here suggested as a likely cause for the relatively warmer marine conditions during MIS 5 compared to MIS 1. The cores also suggest distinct differences between the glacial intervals MIS 2 and MIS 6. While MIS 6 is distinguished by a relatively thick sediment unit poor in foraminifera and with low Mn values, MIS 2 is practically missing. We speculate that this could be the effect from a paleocrystic sea-ice cover north of Greenland during MIS 2 that prevented sediment delivery from sea ice and icebergs. In contrast, the thick sequence deposited during MIS 6 indicates a longer glacial period with dynamic intervals characterized by huge drifting icebergs delivering ice rafted debris (IRD). A drastic shift from thinner sedimentary cycles where interglacial sediment parameters indicate more severe sea-ice conditions gave way to larger amplitude cycles with more open water indicators was observed around the boundary between MIS 7/8. This shift is in agreement with a sedimentary regime shift previously identified in the Eurasian Basin and may be an indicator for the growth of larger ice sheets on the Eurasian landmass during the penultimate glacial period.

1. Introduction

Our understanding of Earth's future climate development is to a large extent based on analogs drawn from previous intervals in the geological history. For example, the conditions prevailing during the interglacial period Marine Isotope Stage (MIS) 11 (Loutre and Berger, 2003), or sometimes MIS 5 (Friddell et al., 2002), are often used as analogs for the future development of our present warm period.

Although it is now clear that orbital forcing exerts the dominating control on Pliocene to Pleistocene glacial cycles (Lourens et al., 2010), it has also been realized that the individual

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interglacials are not directly interchangeable (Droxler et al., 2003), but rather develop in their own unique way (e.g. Candy et al., 2014; Droxler et al., 2003). To a certain extent, the climatic evolution of the individual interglacials is driven and controlled by contingency depending on how factors such as ice-sheet configuration and oceanic circulation developed during the preceding glacial. The preceding glacial in turn was partly determined by conditions during the previous interglacial, and so on. Consequently, the climatic development of individual glacial and interglacial stages during the Pleistocene cannot be understood only as a simple function of geographic boundary conditions and variations in solar radiation received at different latitudes.

In a compilation of ice, marine, and terrestrial records covering the past 800 ky, Lang and Wolff (2011) demonstrated that the last five glacial cycles typically (with the exception of MIS 7) are characterized by strong interglacials and glacials. For example, benthic $\delta^{18}O$ values of the last five interglacials lie within a few tenths permil of each other, while the interglacial-glacial contrast remains stable around 2 permil (Imbrie et al., 1984; Lisiecki and Raymo, 2005). Similarly, temperature and CO_2 records from Antarctic ice cores also show similar levels during the last five interglacials (Loulergue et al., 2008; Luthi et al., 2008). This suggests similar conditions during all five interglacials, although there were differences in the duration and likely also ice sheet configuration.

Mid-latitude terrestrial climate archives, such as loess (An et al., 2014; Vidic et al., 2003) or speleothems (Winograd et al., 1997) show clear glacial-interglacial swings where interglacial conditions appears to have been quite similar from interglacial to interglacial. For example, although the thickness of the interglacial soil horizons in the Chinese loess sequences may vary, proxy parameters such as magnetic susceptibility or quartz grain size display similar values for the last five interglacials, or so. Furthermore, sea surface temperature records from e.g. South Atlantic (Hodell et al., 2000), the tropical Pacific (Lea et al., 2003), and the northern North Atlantic (Bauch and Erlenkeuser, 2003; Kandiano and Bauch, 2003; Lawrence et al., 2009) also show similar conditions during the last five interglacials and glacials, respectively.

Thus, although there certainly were significant differences in how the individual interglacials developed in lower latitudes, they all remain within the boundaries of what can be considered typical interglacial conditions. In contrast, glacial and interglacial conditions in the Arctic Ocean north of Greenland display a considerably larger range of variability (Jakobsson et al., 2010; Nørgaard-Pedersen et al., 2007a). For example, although global ice volume (Lisiecki and Raymo, 2005), sea level (Lambeck et al., 2002; Waelbroeck et al., 2002), greenhouse gas concentrations (Köhler et al., 2010), and solar insolation (Berger and Loutre, 1991) were similar during the last glacial maximum (MIS 2) and the penultimate glacial maximum (MIS 6), shelf and sea-ice conditions appear to have been drastically different between the two glacials in the Arctic (Arndt et al., 2014; de Vernal et al., 2013; Jakobsson et al., 2014, 2010; Knies et al., 2001; Niessen et al., 2013; Polyak et al., 2010). The aim of this study is to compare how the glacials and interglacials developed over the last two glacial cycles in a set of cores from the Morris Jesup Rise and the Lomonosov Ridge north of Greenland. This area is today characterized by the most severe sea-ice conditions in the Arctic Ocean (Comiso, 2011; Comiso et al., 2008; Kwok et al., 2013; Maslowski et al., 2012; Wang and Overland, 2009), and could act as an indicator for the entire Arctic. If the area north of Greenland was sea-ice free, then likely all of the Arctic Ocean was open (Polyak and Jakobsson, 2011). By combining sedimentological, geochemical, and biological data from the various glacial and interglacial stages of the last two glacial cycles, differences in ocean circulation and sea-ice conditions can be assessed (Stein, 2008).

2. Background − sea ice and circulation

Surface circulation controlling the sea-ice drift is dominated by two major patterns, the Transpolar Drift bringing sea ice from the Siberian shelves across the Eurasian Basin to the Fram Strait, and the Beaufort Gyre describing an anticyclonic circulation over much of the Amerasian Basin (Aagaard et al., 1985; Lisitzin, 2002; Meier et al., 2014; Peralta-Ferriz et al., 2013; Pnyushkov et al., 2015; Rudels, 2015). The boundary between the two systems is to a large extent reflected in the composition of the sediment on the sea floor. The Transpolar Drift brings material from the Siberian shelves to the Eurasian basin and effectively limits the input of Laurentian material to the Eurasian Basin. While the Beaufort Gyre primarily distributes material from the Canadian and East Siberian shelf areas, it may also incorporate material from the Transpolar Drift at the boundary between the two systems and redistribute this into the Amerasian Basin (Rudels, 2015; Stein, 2008). Today the position of the boundary is controlled by the mode of the Arctic Oscillation (Kwok et al., 2013; Mysak, 2001; Serreze and Barry, 2014). In an overview of the available sedimentary data, Polyak and Jakobsson (2011) concluded that the present circulation pattern with the Transpolar Drift and the Beaufort Gyre has been characteristic of the Arctic Ocean at least back to 650 ka (see also Cronin et al., 2008).

As a result of the convergence of the Beaufort Gyre and the Transpolar Drift, sea ice tend to accumulate in the Canadian Archipelago and in the area north of Greenland. Areas which today are characterized by some of the most severe sea-ice conditions in the Arctic (Haas et al., 2006; Rothrock et al., 2003; Stein, 2015). Here, densely spaced pressure ridges, counting to the thickest in the Arctic Ocean, make icebreaker expeditions slow and cumbersome, and open waters in the form of polynyas rarely develop even in the summer months (Bourke and Garrett, 1987). Observations on sea-ice variations over the past decades (Comiso and Hall, 2014) and numerical simulations suggest that this area might retain sea ice even if the rest of the Arctic should become ice free in the future (Comiso and Parkinson, 2004; Comiso et al., 2008; Wang and Overland, 2009). Reconstructions of past variations in sea-ice variability, both thickness and coverage, are notoriously difficult both because of the scarcity of suitable proxies and because of the sparse sampling in many parts of the Arctic Ocean. Nevertheless, a few models and reconstructions indicate a possible breakdown of the perennial sea ice with open waters north of Greenland during earlier parts of the Holocene warm period (Funder et al., 2011; Stranne et al., 2014).

Sea ice coverage and thickness are influenced by the inflow of relatively warm North Atlantic intermediate waters (NAIW) into the Arctic basin through the Fram Strait (e.g., Rudels et al., 2012; Steele and Boyd, 1998). However, the processes controlling the exchange with the Atlantic Ocean and the formation of the Arctic Bottom Waters are complex and involve brine rejection on the shelves and exchange of water masses between the different Arctic basins (Aagaard et al., 1985; Arthun et al., 2011; Gordon, 1986; Rudels et al., 2005). The Lomonosov Ridge bisecting the central Arctic Ocean is in this respect a key area. It is here that deep water exchange between the Amerasian and Eurasian basins takes place (Björk et al., 2007), and variations in the vertical position of the different water masses therefore influence the global deep water circulation (Björk et al., 2010; Jakobsson et al., 2010).

During full glacial times with sea level at least 100 m below the present level (Lambeck et al., 2014), the Arctic Ocean was only connected to the world ocean through the Fram Strait, while the Bering Strait and the connections across the Barents Sea and the Canada archipelago were closed (Boer and Nof, 2004; Brigham-Grette, 2013; Fritz et al., 2012; Jakobsson, 2002; Margold et al., 2015). These changes significantly influenced the fresh water