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Evaporative fractionation of $H_2^{18}O$ in the polar ocean and the invisibility of large changes of ice volume and sea level in the Saalian $\delta^{18}O$ proxy records

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Abstract

The oxygen isotope ratio, $\delta^{18}O$, as measured in skeletons of oceanic foraminifera, is a proxy for changes in world glacial ice volume and sea level and is of fundamental importance in the study of past climate change. However, in the Late Saalian glaciation the $\delta^{18}O$ proxy reflects neither a large increase in glacial ice volume from 155 ka-142 ka nor the subsequent major deglaciation from 142 ka-136.5 ka in which sea level rose from about -140 m to +4 m at the end of the Drenthe sub stage. This deglaciation was caused by a large reduction in the ice sheet moisture supply caused by storm path diversion associated with the loss of thermohaline circulation and the capping of the high latitude North Atlantic with melt water that passed through the Mediterranean Sea from the overflow of a giant ice-blocked Siberian lake. The $\delta^{18}O$ proxy also fails to record the subsequent ice volume buildup of the Warthe substage in which sea level fell about 80 m. The explanation for the invisibility involves the fractionation and sequestration of large amounts of $H_2{}^{18}O$ in the effectively isolated and sea-ice-free deep polar ocean. The sequestration and subsequent release of polar water, enriched in $H_2{}^{18}O$, distorted the fractionation record in the world ocean and destroyed the accuracy of the $\delta^{18}O$ proxy. The proposed physical consequences of the ice-free polar ocean also include a sea-ice-free Labrador Sea and Baffin Bay and an oceanic circulation mode that drove the Drenthe sub stage glaciation to its maximum extent in Eurasia. The initial cause for the anomalous effects in the $\delta^{18}O$ record was the known ice-flow blockage of all the Siberian rivers that discharge into the polar ocean west of the Lena River. Without adequate stratification by inflowing fresh river water, sea ice was unable to freeze on the deep polar ocean and the fractionation and the physical changes in the polar ocean that are described here followed.

Introduction

Arkhipov et al, (1995) published a summary of glacial drainage conditions at the maxima of the Weichselian and Saalian glaciations that occurred in the last two ice age cycles beginning about 160 ka (ka = thousands of years before present). This paper was of fundamental interest because it also outlined a concept of major deglaciation caused by the reduction of precipitation on the ice sheets by means of atmospheric circulation changes associated with changes in thermohaline oceanic circulation. The deglaciation of the Drenthe sub stage of the Saalian glaciation was the cited example. However, this concept was contrary to the popular Milankovitch hypothesis (Milankovitch, 1930; Hodell, 2016) of direct control of glacial changes by Northern Hemisphere orbital high latitude summer insolation, a concept that has been supported by frequency correlations in the $\delta^{18}O$ records (Hays et al., 1976). Furthermore, there was no support for a Drenthe substage deglaciation in the oceanic oxygen isotope ratio data. This contrasts with the deglaciation of the following Warthe substage that is easily identified with Termination 2 in the $\delta^{18}O$ records.

Since 1995 the published results of the analysis of sea stands at locations having varying uplift rates on Barbados (Johnson, 2001) have established a sea level slightly above present at 136.5 ka that is identified with the Drenthe deglaciation. In addition the causative aspect of the simple Milankovitch hypothesis has been weakened (Wunsch, 2004) who showed statistically that the variance in the $\delta^{18}O$ sea level proxy is largely

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random over long time intervals. In the data discussed in this paper, the observed sea level rise that was caused by the deglaciation of the Drenthe substage during an insolation minimum, and the sea level fall due to the Warthe glacial ice accumulation under steadily rising Northern summer insolation are both in conflict with the Milankovitch hypothesis, and are invisible in the oceanic δ^{18}O record. This invisibility is a consequence of fractionation in the ice-free polar ocean. However, to grasp the complete significance of the fractionation it is necessary to examine its cause, and the associated effects of the cause on the strength and location of deep-water formation, that is, on the thermohaline circulation in the northern North Atlantic. Therefore in this paper Sects. 1-4 describe the physical principles and evidence for events that led to the ice-free polar ocean and the fractionation. Sects. 5-7 outline the sequence of fractionation events and others that resulted from the ice-free polar ocean.

1 Defining the invisibility problem and the related importance of storm path location

1.1 The δ^{18}O oxygen isotope ratio proxy for world sea level and conflicting coral reef data

The riddle of the cyclic Plio-Pleistocene ice ages with their wide swings of world sea level and ice volume has been largely defined by δ^{18}O, the algebraic form of the oxygen isotope ratio proxy for changes in glacial ice volume and world sea level. As measured in continuously deposited deep-sea sediments that are rich in foraminifera, the δ^{18}O isotope ratio has long been a useful tool in the investigations of paleoclimates. The changes in the ratio occur because the heavier molecule of ocean water, $H_2{}^{18}O$, evaporates much more slowly than $H_2{}^{16}O$ thus fractionating and enriching $H_2{}^{18}O$ in the world ocean when the depleted precipitation is locked into glacial ice. Ideally, the measured δ^{18}O is a good proxy for changes in glacial ice volume and the resulting variations in world sea level, and is particularly relevant in attempts to determine the cause of ice age climate variations.

Wunsch (2004) has shown by extensive analysis that most of the δ^{18}O proxy variance is not explained by high latitude insolation changes, and is more likely to have been caused by internal variations within Earth's climate system. Therefore despite the correlations (Hays et al., 1976), the Milankovitch concept of orbital control of high latitude Northern summer insolation can not be a significant cause of the glacial climate variations. To identify the dominant causes it is important to find out how the internal factors act with the external orbital effects to cause the observed δ^{18}O proxy changes. This paper is an effort to contribute to that goal by explaining the conflict between an inferred low sea level found in the δ^{18}O proxy records of deep-sea cores during the Drenthe substage of the Saalian glaciation (Fig. 1), and the physical evidence for a simultaneous paleo sea stand at 4 m above present at 136.5 ka. A related conflict occurs in the subsequent Warthe substage of the Saalian. The Saalian name is used here for the entire Eurasian glaciation of that time. The explanation for this conflict involves the blockage of Siberian rivers by glacial ice flow, a quite different ice-free polar ocean circulation, and the fractionation of $H_2{}^{18}O$ and its sequestration in the deep polar ocean. The conflict between observed sea level change and the isotope ratio proxy ended after orbitally controlled summer insolation (Berger, 1978) generated very strong African monsoons (Rossignol-Strick, 1983; Rossignal-Strick et al., 1998) that capped the northern North Atlantic with lower salinity Mediterranean outflow and melt water and started the Warthe deglaciation of Termination 2.

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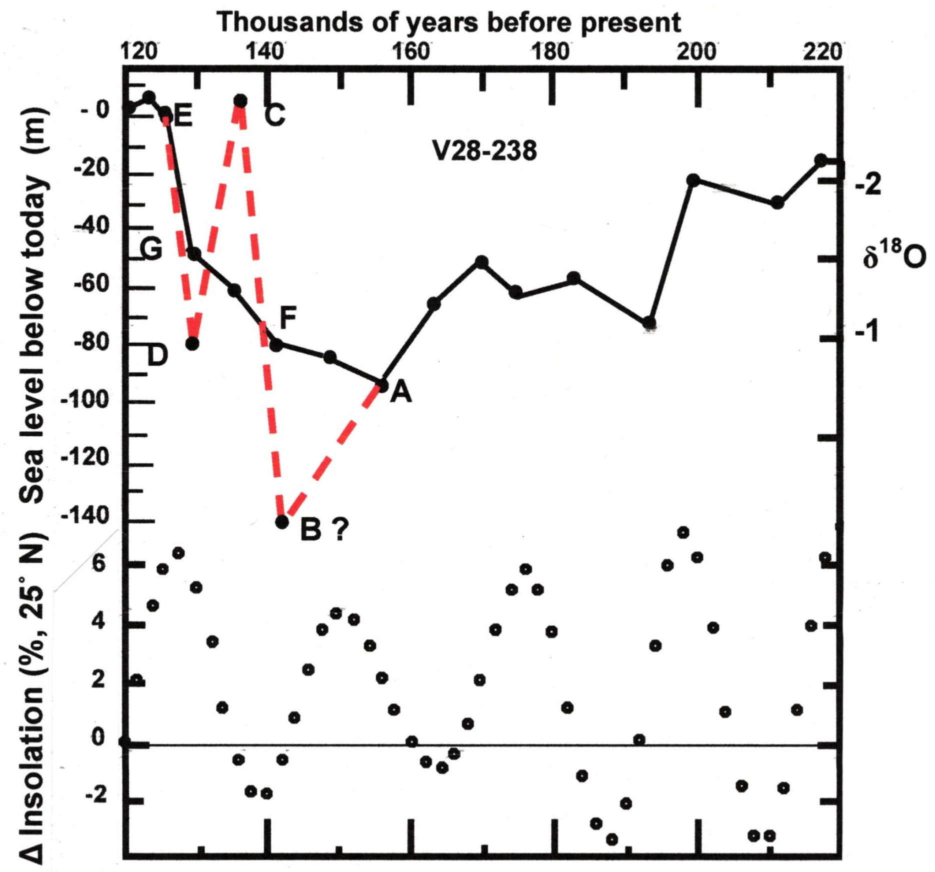

Figure 1 Oxygen isotope ratio record for a benthic species in core V28-238 during the Saalian glaciation from 200 ka to 126 ka. Dashed red line is the proposed true world sea level after the blockage of Siberian rivers by glacial ice began about 156 ka. Dotted cycles are % deviation of the Northern caloric summer insolation average at 25° N from the modern 1950 value of 918 langleys day^{-1} (3.84x10^{7} jm^{-2}day^{-1}). δ^{18}O values from Shackleton and Opdyke (1973). Insolation from tables supplied by Berger (1978).

The Drenthe high sea stand anomaly has been known for some time from prominent fossil coral reefs of controversial age on the rapidly uplifted Huon Peninsula in New Guinea (Chappell,1974). The same sea stand is found on the more slowly uplifted island of Barbados where most of the resulting reef structure, except for the wave-cut erosion at the brief maximum sea stand itself, is buried beneath the subsequent coral deposits of the last interglacial. This sea stand on Barbados has been dated by a differential uplift method (Johnson, 2001) that is grounded on reliable thorium-uranium dates for Australian corals that grew at the end of the last interglacial (Stirling et al., 1995) and which, after a small correction (Johnson, 2015a) yields a sea level of +4m at 136.5 ka. This deglaciation occurred during an

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interval of low Northern summer insolation, and it is diametrically opposed to the hypothesis (Milankovitch, 1930) of control of glaciation by orbital effects on Northern Hemisphere high latitude solar insolation

1.2 Storm path variation: The importance of storm path location

1.2.1 A high Arctic example

To interpret the complex data from Earth's climate system during the Saalian glaciation it is necessary to consider the critical conditions within the system that change precipitation rates on the world ice sheets. Although cold climate temperatures are necessary for the maintenance of an ice sheet, the moisture supply is equally important. The rate of growth or shrinkage of an ice sheet depends on the balance between loss of ice due to seasonal melting or discharge into the sea and the gain of ice due to storm precipitation. Without adequate moisture supplied by storms, ice sheets shrink. With a good supply they grow. On the matter of world glacial ice volume growth and shrinkage, the importance of the paths of storms that carry moisture from the ocean to the ice sheets cannot be overstated. Obvious examples are arid Peary Land on the northern tip of Greenland and much of Ellesmere Island to the west. The climate is quite cold in these regions, with Peary Land only 800 km from the pole. Yet these areas are largely free of glacial ice because storms are infrequent and precipitation rates today are quite low. But the coasts are deeply indented by fiords, indicating that in times past large ice sheets, fed by frequent storms, occupied the land, see Sect. 5.3. Now the closest frequent storms are far away to the south. Today, with strong thermohaline circulation (THC) and resulting northward heat transport, storms are steered northeastward through the Greenland-Norwegian Sea basin by the temperature gradient between the cold Greenland ice sheet and the warmer water of the Greenland-Norwegian Sea, and much of their precipitation falls into the ocean. In most ice ages with a modest reduction of today's interglacial THC, the sea surface in the basin cools. Storm paths then shift southward and pass over Scandinavia, feeding the ice sheet there.

1.2.2 A modern example: the Little Ice Age cycle

As seen in the context of the 1500 yr deep ocean oscillation (Johnson, 2015b), the sea surface salinity (SSS) of the Greenland-Norwegian Sea basin plays a role in the THC, and the Little Ice Age can be explained by the effect of the varying basin salinity on the deep circulation and on the sea surface temperature (SST) gradients. The paths of storms tend to lie perpendicular to the strongest gradients. When the basin SSS is high, a greater amount of sinking deep-water feeds the large-scale THC. The sinking water is then replaced by a stronger North Atlantic Drift, the basin SST is warmer, and storms extend far northward in the Greenland-Norwegian Sea. When the basin SSS is lower, the THC is weaker, the basin SST is also lower, and the stronger gradient and storm paths lie to the south, with more storms extending over Scandinavia. The SSS would have been near a maximum a thousand years ago when populations were large in the cooler northern parts of the British Isles, and the pastoral Norse established colonies on southern Greenland. But as the basin salinity was reduced by the deep ocean oscillation, the THC became weaker, the climate cooled on Greenland and the British Isles, and those higher latitude areas were largely abandoned. Near the Little Ice Age extreme about the year 1750, as described by Lamb, wine grapes no longer grew in southern England and the winter sea ice in some years extended almost as far south as northern Scotland (Lamb, 1995).

In contrast, the salinity in the basin is today increasing because of the cyclic deep ocean oscillation, and also because of the effect of modern society's technical activities on the Mediterranean Sea salinity. Excess atmospheric CO_2 warming should increase evaporation rates in the arid zones that include the Mediterranean Sea. The rising evaporation rates and the dams constructed for irrigation on nearly all the major rivers discharging into the Mediterranean are probably causing the measured increase in Mediterranean salinity and that of its outflow, which feeds significant saltier water northward into the Greenland-Norwegian Sea basin. Before these modern technical effects appeared, the cyclic salinity increase in the basin was already limiting the southward extent of the mid winter sea ice to the southern tip of Spitsbergen in most winters (Lamb, 1972). Now however, with technically enhanced salinity, the warmer southern water replacing the greater amount of sinking surface water in the THC keeps the mid winter sea ice limit well to the north of Spitsbergen's north coast, as shown by satellite images posted on the internet by the Technical University of Denmark in recent years. The lesson we get from the Little Ice Age's 1500 yr cycle is that the sea surface salinity in the Greenland-Norwegian sea basin determines the paths of the storms, and if the SSS becomes extremely low deep water would not form there and storm paths would be much farther to the south than society has seen in historic times.

1.3 A long term orbital factor: orbitally-controlled African monsoons

In the matter of shifting northern North Atlantic storm paths, African monsoons can play an indirect but decisive role. Strong monsoons occur when Earth in its eccentric orbit comes closest to the sun during Northern summer due to the 23,000 yr cyclic precession effect. Higher summer temperatures in the large land area of northern Africa relative to temperatures in the ocean areas and smaller African land area south of the equator then generate quite strong monsoons, which can extend northward over the Sahara and greatly increase Nile River flow (Rossignol-Strick, 1983; Rossignal-Strick et al., 1998). Strong monsoons therefore reduce the salinity of the Mediterranean Sea and its outflow at Gibraltar and the salinity of the higher latitudes where winter-cooled saline surface water sinks, thus driving the THC. The mixing time of the Mediterranean Sea is about 120 yrs as estimated from the exchange current flow (Bryden and Kinder, 1991) and the intervals of strong monsoons can last for ~4,000 yrs. Consequently the Mediterranean's reduced salinity and outflow closely tracks the occurrence of the strong monsoons. If not inhibited by a cold glacial climate extreme in northern Africa, the strong monsoons can reduce the THC, cool the high latitudes, and move the storm paths far to the south away from the Canadian and Scandinavian ice sheets giving them a dry glacial climate. On the other hand, warmer water in the sub polar North Atlantic adjacent to cold land tends to bring heavy precipitation over the land and cause ice sheet growth as reported by Ruddiman and McIntyre (1979). A somewhat different mechanism affecting Eurasian glaciation probably occurred due to sea-land temperature contrasts in western Europe and northwest Africa, as discussed in Sect. 4.2.

The orbital-monsoon mechanism is important, but like the hypothetical Milankovitch mechanism of direct control of glaciation by high latitude insolation, it can likewise be overcome by other effects within the climate system. An important example is framed in the following question: Why is decreasing insolation so often associated with glacial growth in the isotope ratio records, whereas the same level of increasing insolation seemingly causes deglaciation? Again the answer is found in the paths of storms governed by temperature gradients over the high latitude ocean. Weak and falling insolation at latitude 25˚ N does cause glacial ice volume to increase by weakening the monsoons, which then increases high-latitude North Atlantic salinity and THC during already-established glaciation, and so indirectly steers storms over growing ice sheets. But at the same level of rising insolation massive Northern ice sheets are more likely to have been accumulated, and any factor that increases their melt water output may further decrease the high latitude salinity, cool the northern sea surface, and move the high latitude North Atlantic storm paths away from the ice sheets and far to the south, thus starving them of moisture. This is particularly true for the Canadian and Scandinavian ice sheets because their melt water is discharged into the areas of sinking THC water and this tends to reduce salinity and sinking rate and maintain colder sea surfaces. Other events also do so, including ice dam failures in Hudson Strait (Johnson and Lauritzen, 1995) and in the Baltic Sea outlet, and the Heinrich iceberg events (Heinrich, 1988), that is, the "collapse" of the great Canadian ice sheet. In the δ^{18}O records of the Pleistocene it is only when the large ice volumes are inferred at glacial extremes that major ice age terminations occur. It is in just such a context of the Saalian glacial extreme that the overflow of a giant ice-blocked lake in Siberia (Fig. 2) eliminated North Atlantic high latitude THC, and triggered a dry-climate deglaciation that brought the world ocean up to a sea level of +4 m with only a small change in the δ^{18}O record (Johnson, 2001; Johnson, 2015a.

2 Where the δ^{18}O proxy record failed: inferred Late Saalian ice volume and sea level changes in the benthic sediment record of core V28-238

Core V28-238 (Fig. 1) was recovered from a depth of 3120 m in the western Pacific where sedimentation rates are low and uniform (Shackleton and Opdyke, 1973). Benthic foraminifera living at that location and depth experienced a relatively constant temperature close to 0° C. Therefore the δ^{18}O measured in their skeletons should provide a good record of ice volume and world sea level changes over time. This is approximately true because δ^{18}O values from that core and core V28-239 were used with calculated insolation variations (Berger, 1978) to obtain the first correct date of about 790 ka for earth's last magnetic field reversal (Johnson, 1982). Extensive statistical analysis of the records of these two and many other cores has shown that the δ^{18}O changes have the same principal frequencies as variations in orbital insolation that are caused by the precession effect, which determines the position of summer in the earth's orbit, and by small changes in the tilt of the earth's polar axis (Hays et al., 1976). Nevertheless, the core records are not free of proxy error, as in the V28-238 record of the Saalian glaciation discussed here. In this core a puzzling departure from the statistical expectation occurs at 175 ka when a somewhat lower (more positive) δ^{18}O value is found instead of a negative peak consistent with higher insolation and stronger monsoons. At 150 ka there is no δ^{18}O evidence for deglaciation despite an insolation peak (Berger, 1978) at 25° N that is almost identical to the peak at 11 ka that is associated with the deglaciation of Termination 1 near the beginning of the Holocene. A third example is the focus of this paper, which is the difference between the observed and dated high sea stand at 136.5 ka (Johnson, 2001; Johnson, 2015a) and the more positive values (lower on the page in conventional plots) of all the world ocean δ^{18}O records during that deglaciation, which suggest incorrectly that little or no deglaciation occurred during the insolation minimum. In the determination of the amplitude of the δ^{18}O proxy changes, the internal factors within Earth's climate system are apparently stronger than the direct effects of insolation variations, such as changes in glacial melting rates. This is emphatically true for the deglaciation of the Drenthe sub stage of the Saalian glaciation.

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3 Why the $\delta^{18}O$ proxy record failed: the blockage of Siberian rivers by glacial ice flow

3.1 The Eurasian ice sheet and its giant lake

The largest Eurasian ice sheet of the Pleistocene may have occurred at the extreme of the Saalian glaciation (Arkhipov et al.,1995), depicted in Figure 2. Glacial ice covered most of the British Isles and extended without a break across Scandinavia and western Siberia almost to the Lena River. To the north it covered the shallow Barents Sea and the Svalbard archipeligo. To the south it reached into the Dniepr River valley not far from the Black Sea. After a weak interstadial about 170 ka, the falling (less negative) $\delta^{18}O$ values in Figure 1 indicate continuing ice sheet growth to point **A** in Figure 1 at 156 ka. The positive fall of the $\delta^{18}O$ then stopped because polar ocean fractionation began, as discussed in Sect. 5. But Drenthe glacial ice continued to accumulate with profound

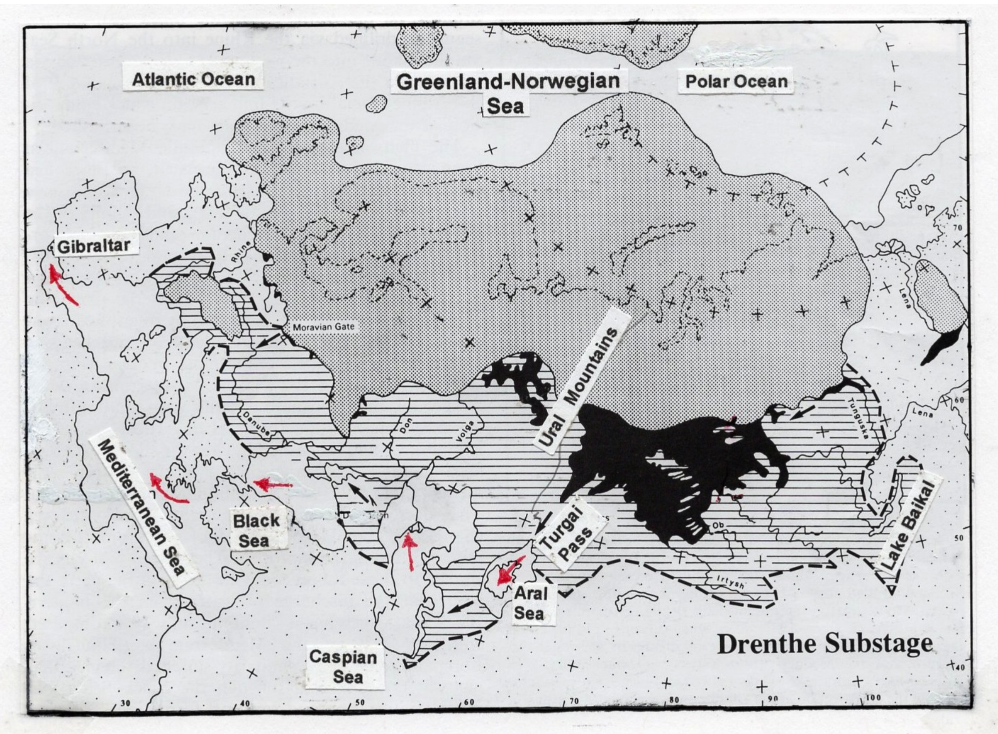

Figure 2 **The maximum extent of the Drenthe substage of the Saalian glaciation at 142 ka. Cross-hatched area is the catchment for melt water and precipitation that drained through the Mediterranean Sea to the North Atlantic during the Drenthe deglaciation. Slightly modified from Arkhipov et al. (1995).**

consequences, and sea level continued to fall. Beginning probably a little before 155 ka in the latter part of the Saalian cycle with the Mackenzie River covered by the Canadian ice sheet of that time, the two largest Siberian rivers, the Ob, the Yenisey, and other smaller rivers that also discharge into the polar ocean became blocked by glacial ice flow. A third major Siberian river, the Lena, remained unblocked, but in the dry glacial climate its discharge into the polar ocean would have been small. Consequently the deep polar ocean was no longer effectively capped and stratified by river water. Sea ice there would have been largely absent throughout the year because in unstratified deep water vertical convection would prevent cooling to the freezing point. As the polar ice disappeared, all the polar changes discussed in Sect. 5 began to occur, while accompanied by the further buildup of world glacial ice and the rise of the water level of the giant lake in Siberia.

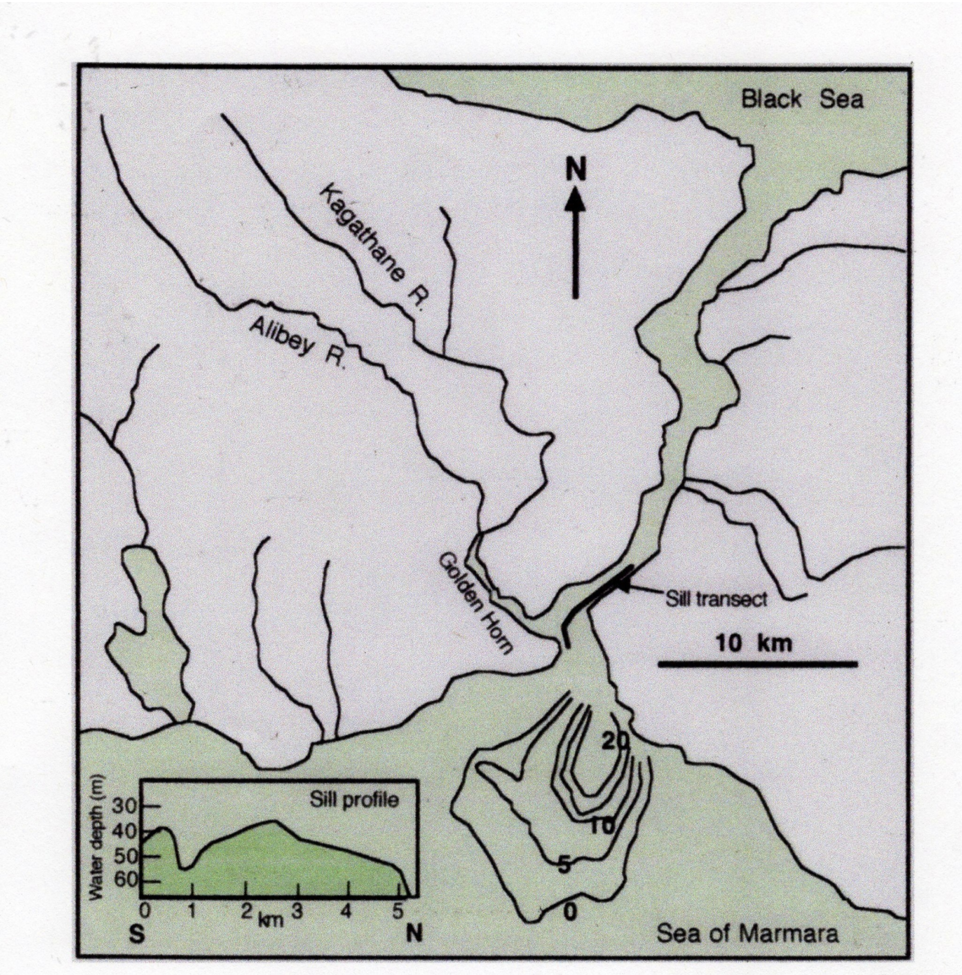

Figure 3 **The Bosporus Strait with sill depth profile. North of the sill transect (heavy dark line) water depths are 90 m-100 m. Minimum bedrock depth at the sill is 65 m below present sea level. Minimum sediment depth is 35 m. The delta profiles are sonar reflection times that depict sediments discharged through the Golden Horn in more pluvial times. Combined from Gunnerson and Özturgut (1974) and Alavi et al. (1989).**

The catchment area of the ice-blocked Siberian rivers extended 3000 km from the Ural Mountains to Lake Baikal (Arkhipov et al., 1995). Within that area the ice-blocked lake began to form. The exact date of the start of the blockage is unknown, but in the dry glacial climate the lake level slowly rose over thousands of years to a maximum and began to overflow through the Turgai pass at its present elevation of 121 m (Google Earth). At that maximum the lake area was about 1,100,00 km^2, or three times the area of the modern Caspian Sea.

The overflow would have begun as a somewhat abrupt event about 142 ka if, as is likely, rapid erosion of sediment in the pass occurred. The overflow quickly filled the basins of the Aral, Caspian, and Black Seas with mixed melt water that flowed onward through the Bosporus strait (Fig. 3), through the Mediterranean, and into the North Atlantic at Gibraltar. In the arid glacial climate before the overflow, the water levels in those isolated inland seas would have been quite low with limited river inputs.

3.2 A pulse of melt water from the Black Sea: the Bosporus sediment dam

A strong pulse of giant lake melt water into the North Atlantic would have been enabled by a sediment dam failure at the south end of the Bosporus. The pulse capped the high latitude North Atlantic with lower salinity water and initiated deglaciation. A Holocene sediment dam analog is known. At a calendar age of 9.4 ka (Major et al., 2006) a rising world sea level at about -20 m washed out a sediment dam on the -65 m bedrock sill (Fig. 3) and flooded northward into the arid Black Sea basin, with the Black Sea level initially down at -100 m or more. The ~40 m of sediment on the sill had been deposited during a high sea level at a more pluvial time like the present when the Bosporus flow was mainly out of the Black Sea into the Mediterranean.

At high interglacial or interstadial sea levels the Bosporus flow is a two-way exchange flow, but as ice sheets grow and the world sea level falls, the flow would become only outward into the Mediterranean until cut off by the increasingly dry glacial climate in the Black Sea catchment. During pluvial interglacial times the sediment from the Kagathane and Alibey Rivers is carried into the strait through the Golden Horn at the south end of the bedrock sill (Fig. 3). Before the cutoff, the southward flow of the Bosporus current over the sill entrains Mediterranean water, which is replaced by a northward bottom inflow that moves the sediment up onto the sill. The present sediment thickness is about 30 m, and repetitive dams of at least that thickness would be expected after times of previous cyclic sea level maxima in the Pleistocene. The giant lake overflow and resulting events are supported by an eastern Mediterranean deep-sea sediment record.

4 The Mediterranean record of the chain of Late Saalian events that were associated with the failure of the world ice volume $\delta^{18}O$ proxy

4.1 Dating the Mediterranean record of core TR172-22

Evidence of the giant lake overflow and subsequent events are found in sediments of eastern Mediterranean core TR172-22 as reported by Thunell and Williams (1982). This core was recovered from a site half way between Crete and Cyprus. The core data replotted in Figure 4 are dated by centering the low-salinity interval of absent foraminiferal data on the accurately calculated 25˚ N caloric summer insolation maximum at 128 ka (Berger, 1978) when the strongest monsoons occurred and Mediterranean salinity was at a minimum. Uniform sedimentation is assumed. Beginning at 130 ka the swollen Nile River made the eastern Mediterranean surface water too fresh

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for salt-water foraminifera to survive. The strong monsoons and large Nile flow imply an absence of summer cooling in northern Africa due to the less extreme extent of the ice sheets at the end of the Warthe substage of the Saalian, about 130 ka. This contrasts with the glacial cooling of northern Africa that apparently occurred toward the extreme of the Drenthe sub stage about 150 ka despite high insolation (Fig. 1). The cooling probably prevented strong monsoons and any deglacial $\delta^{18}O$ proxy response at that time. Such a strong cooling may have caused a glaciation in the Atlas Mountains of northwestern Africa (Hughes et al., 2011).

4.2 The Mediterranean record: The giant lake overflow

At or shortly before 142 ka when world sea level was quite low, the giant Siberian lake burst through sediment constraints at the Turgai pass, successively flooded the isolated Aral and Caspian basins, and poured into the Black Sea through the Manytsh channel (Arkhipov et al., 1995). When the overflow at the Bosporus began, its sediment dam (Fig. 3) was abruptly washed out and the Black Sea level would have fallen rapidly

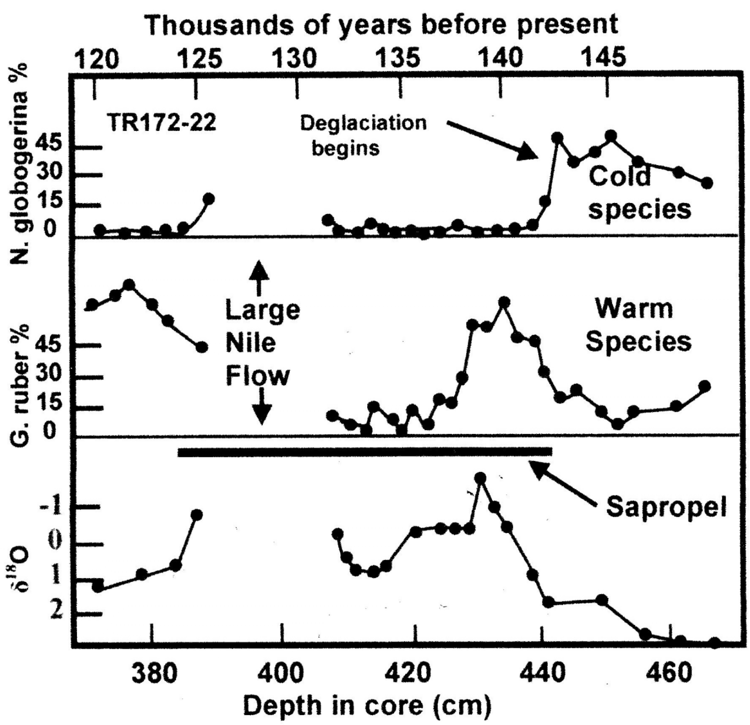

Figure 4 Sedimentary records from core TR172-22 reported by Thunell and Williams (1982) in the eastern Mediterranean from the beginning of the Drenthe deglaciation to the end of Termination 2. The maximum strength of African monsoons and minimum Mediterranean salinity coincides with the astronomically calculated summer insolation maximum at latitude 25° N. Therefore the time scale is set with the center of the interval of absent salt water foraminifera and maximum Nile River flow at 128 ka.

by possibly as much as 40 m to the present bedrock sill elevation of -65 m. That event is dated at 142 ka by melt water stratification that preserved summer warmth in the mixed surface layer of the Mediterranean. This caused the disappearance of the cold-water foraminiferal *N. globogerina* (top, Fig. 4), and increased the abundance of the warm water species, *G. ruber,* (middle, Fig. 4). Before the flooding event the Black Sea could have had salinity like today's, about 18 ‰. After the mixing that occurred during the flooding, the Black Sea surface salinity would have been somewhat less, possibly less than half the Mediterranean salinity. Therefore the stratification in the Mediterranean was quite effective and would have limited the convective circulation that usually oxidizes benthic sediments, thus initiating the formation of the black sapropel sediments in the eastern Mediterranean (Fig. 4).

If the Bosporus sediment on the sill had been only 30 m thick as it is today, the rapid washing out of the dam would have put a large pulse of mixed low-salinity water equivalent to a 3.2 m layer of the melt water on the Mediterranean surface. This would have injected a corresponding pulse of lower salinity surface water into the North Atlantic at Gibraltar. The outflow pulse apparently capped the high latitude seas with water of reduced salinity that shut down THC, thus reducing North Atlantic Drift flow from the Gulf Stream and strengthening the southward return flow of the Gulf Stream in the eastern North Atlantic.

4.3 The Drenthe deglaciation

The loss of the THC would have initiated a deglaciation in Antarctica as well as North America and Eurasia. With the lack of North Atlantic thermohaline circulation and resulting loss of mid level inputs of warmer deep-water to the Southern Ocean, the seasonal winter sea ice extent around the Antarctic continent would always been at a maximum (Johnson, 2015b). Therefore circumpolar winter storms, moving perpendicular to the strongest land-sea temperature gradient, remained more distant from the continent with a severe reduction in Antarctic continental precipitation. In Europe and Asia, it was probably another seasonal shift in storm paths that rapidly diminished the Saalian ice sheet. The perennially colder seas bordering Europe to the west would not have been as cold as the iceberg-filled seas during Heinrich events, but would still have caused a significant increase in the land-sea temperature gradients there, due to the reduced North Atlantic Drift and the resulting more sharply defined colder return flow of the Gulf Stream off the west coast of Spain. Land temperatures in Western Europe in winter were likewise cold, and the jet stream that guided winter storms would have been located on a west-to-east path at the latitude of Spain and well south of the ice, thus preventing the Eurasian ice sheet from receiving heavy snows. But in summer the jet stream would have been quite different. With the summer warming of the land masses, the colder sea–warm land temperature contrasts in France and Spain would have shifted the jet stream to a southwest-northeast path, and the jet stream flow and the storms that follow it would have carried warmer and dryer air from as far south as northwestern Africa into the Eurasian ice sheet region, with a strong deglacial effect.

4.4 A rapid Warthe glaciation terminated by strong monsoons

The Drenthe deglaciation supplied an increasing amount of melt water to the Mediterranean over several thousand years, as suggested by the increasing abundance of warm water foraminifera as stratification became more effective and sea surface temperatures rose (Fig. 4). But when the Siberian rivers again drained into the polar ocean at 136.5 ka, the Turgai pass overflow stopped, and the polar ocean became briefly stratified. However, as the Warthe glacial sub stage began, glacial growth was renewed and the Siberian rivers became quickly blocked again by ice (Arkhipov et al., 1995), although with the shorter duration of the Warthe no overflow of the giant lake occurred. At the start of the Warthe glaciation the initially low insolation and weak monsoons like today increased Mediterranean salinity and high latitude THC. Therefore starting with large residual Northern ice sheets, glacial growth was rapid and sea level fell to about -80 m at 130 ka while summer insolation increased by about 6 % in Northern latitudes. This sea level fall (**D**, Fig. 1) was indicated by dated fossil corals from rapidly uplifted New Guinea as reported by Esat et al. (1999). But strong monsoons began about 130 ka thus increasing the Nile flow, lowering Mediterranean salinity, eliminating the saltwater foraminifera (Fig. 4), and lowering the rate and salinity of the Mediterranean outflow. This reduced the glacial climate THC, altered the storm paths, and initiated the rapid deglaciation of Termination 2 that ended about 126 ka (**E**, Fig. 1), paradoxically under the influence of stronger THC in the northern Greenland Sea basin as argued in the Sect. 6 summary.

5 Why the oceanic $\delta^{18}O$ proxy record failed: fractionation in the polar ocean

5.1 The cause of the anomalous $\delta^{18}O$ record of the late Saalian glaciation

The $\delta^{18}O$ enrichment of the world ocean by evaporative fractionation is the basis for the $\delta^{18}O$ proxy for ice volume and sea level change. World ocean $\delta^{18}O$ values change in a positive direction as glacial ice volume increases. But if an isolated body of water becomes enriched in $H_2{}^{18}O$ by fractionation, the world ocean becomes correspondingly depleted in $H_2{}^{18}O$, making the measured oceanic $\delta^{18}O$ increasingly negative (too high on the page in conventional data plots as in Figure 1). Such a negative error might have had contributions due to fractionation in the isolated Black, Caspian, and Aral seas, or the giant ice-blocked Siberian lake east of the Ural Mountains. However, these inland seas receive water from precipitation that is quite deficient in $H_2{}^{18}O$, and relative to the world ocean their surface areas are small. Therefore their contributions to the $\delta^{18}O$ errors were negligible. The Mediterranean Sea with its vigorous exchange currents at Gibraltar (Bryden and Kinder, 1991) and a present mixing time of about 120 yrs is not sufficiently isolated.

On the other hand the polar ocean (Fig. 5) is large enough and isolated enough at the glacial extreme to have undergone the fractionation and sequestration of $H_2{}^{18}O$ that distorted the world $\delta^{18}O$ record. It was much more isolated than it is today, with the Beringia land bridge exposed and the Barents Sea ice dome blocking flow north of Scandinavia. With lower sea level later in the glacial cycle, the main concern here is the deeper polar ocean. The deep ocean begins at the edge of the continental shelves, taken here to be close to a present depth of 200 m. Depths of 4000 m or more are found in parts of the polar basin. The area with present depths greater than 200 m (Fig. 5) is 5.1×10^6 km^2, twice the area of the Mediterranean Sea or 6.7 million km^2 when the deep Greenland-Norwegian basin is included. The polar ocean's connection to the world ocean was through the 3700 m-deep channel in the Fram Strait into the Greenland-Norwegian basin, and then over the Iceland-Faroe ridge, a broad plateau with present sill depth of about 480 m.

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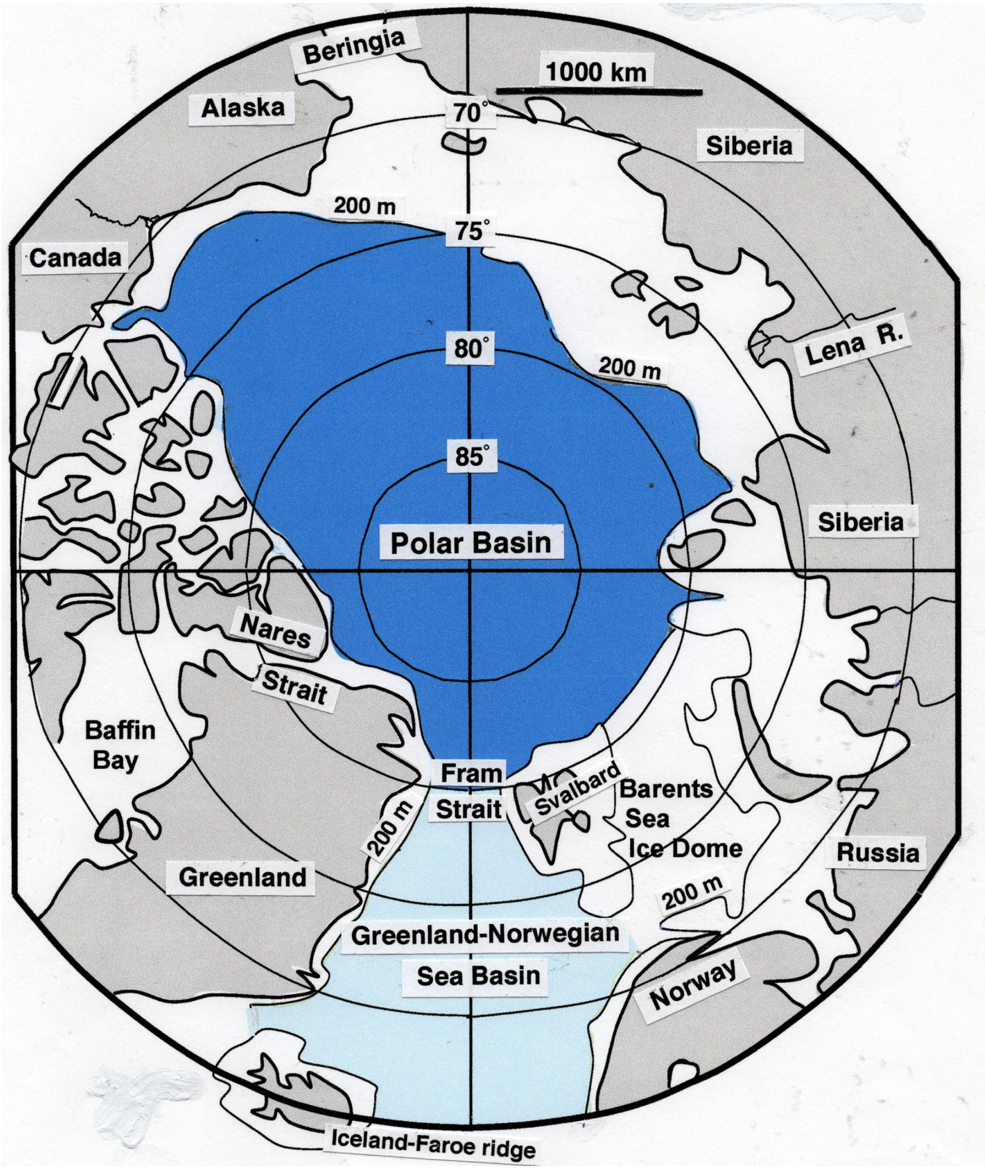

Figure 5 **A sketch map of the North Polar regions using a polar projection to show the deeper parts of the polar ocean basin and the Greenland-Norwegian Sea basin at present depths greater than 200 m. Redrawn from Bartholomew's Advanced Atlas of Modern Geography, 1950, McGraw-Hill Inc., New York.**

The deep polar ocean area of 5.1×10^{6} km^2 is far less than the world ocean area of 320×10^{6} km^2. Nevertheless over the long intervals of time similar amounts of fractionation could have occurred. In Figure 1 in the interval **A-B** an additional world sea level fall of about 50 m probably occurred, which implies that 16×10^{6} km^3 of water were evaporated and precipitated on ice sheets in about 13,000 yrs. The same quantity of polar ocean water would have been evaporated and carried away during that interval if the annual rate of evaporative loss was equivalent to a plausible polar ocean layer 22 cm-thick. From thermodynamic considerations fractionation is more effective for water evaporated at low polar temperatures than for water evaporated in warmer latitudes. Consequently, the low polar temperatures would have contributed to the deficit of $H_2{}^{18}O$ in the world ocean that was slightly larger than the excess caused by additional world ice accumulation, as suggested by the small negative rise of the $\delta^{18}O$ value in Figure 1 between 156 ka and 142 ka.

5.2 Circulation in a sea-ice-free polar ocean

The change to an ice-free polar ocean would have caused important polar circulation differences relative to today. With modern winter sea ice covering the polar ocean, I observed satellite images in year 2011 posted by the Technical University of Denmark on the internet that showed sea ice moving roughly clockwise around the pole, consistent with wind stresses due higher atmospheric pressures over the cold ice. As the flow approached Greenland it split apart with much of the flow going westward, but a minor part continued southward along the eastern Greenland coast. The ice-filled East Greenland Current is therefore now a visible drain on the surface water of the polar basin. In other times like today, the THC is likewise a drain on the Greenland-Norwegian Sea basin.

However, during the long winters near the Drenthe extreme with an open polar ocean, the polar atmosphere would have been warmer than that over surrounding cold lands. The probable result was the generation of large counterclockwise cyclonic storm systems that were dragged slowly eastward by the effect of Earth's rotation on the bulk of the each storm area. The storm centers of rotation were probably well off shore, and the net effect of wind stresses on the sea surface would have driven the oceanic circulation around the polar basin in a counterclockwise direction. The northward projection of the Greenland coast in Figure 5 would therefore not have diverted polar water flow into the East Greenland Current as it does today with clockwise polar flow. If so, the main transfer through the upper levels of the Fram Strait may have been a slow northward drift of water from the adjacent Greenland basin to replace the net polar ocean evaporation losses.

5.3 An explanation for the massive Eurasian glacial ice buildup.

Another consequence of an open polar ocean is a chain of circulation effects that may explain the record-breaking extent of the Eurasian ice sheet during the Saalian glaciation. Northward flow through the Nares Strait between Greenland and Ellesmere Island (Fig. 5) into the polar ocean may have been indirectly responsible for steering winter storms through southern Scandinavia into the heart of the Eurasian ice sheet. Today's flow through the Nares Strait is driven by atmospheric pressure differences over the 500 km length of the strait. In the year 2011 and early 2012, I observed movement of clumps of sea ice in high-resolution satellite images posted by the Technical University of Denmark. On 134 days when flow direction in the strait could be observed, 71 % of the flow-days were north-to-south and 29 % were south-to-north. With only three exceptions, atmospheric pressure differences determined the direction of flow. In 2011 pack ice covered the area between the pole and the Canadian-Greenland northern coasts. But when that area was free of sea ice in the Later Saalian, the warmer atmosphere and

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lower polar atmospheric pressure may have ensured a dominant northward flow from Baffin Bay into the polar ocean, both through the strait and through any open channels in the Queen Elizabeth islands to the west. Therefore without the southward flow of lower density polar water entering and stratifying Baffin Bay, strong land-sea temperature contrasts around the perennially open water in the bay would have generated a powerful low pressure system over Baffin Bay and the Labrador Sea like the one that initiated the last ice age at 120 ka (Johnson, 2015a).

This Labrador Sea Low would then have prevented the development of the nearby Iceland Low cyclonic system that provides prevailing wind stresses that would be quite important for driving saline North Atlantic Drift water northeastward into the Norwegian Current. Today this transport strongly favors thermohaline warming effects west of Norway and north of the Iceland-Faroe ridge (Johnson, 2015a). Consequently in the Late Saalian while the Labrador Sea and polar ocean remained ice-free, Northern Hemisphere thermohaline circulation was likely reduced, but not eliminated because other deep water would have been forming in the north end of the unstratified ice-free Baffin Bay. But without the wind stresses of the Iceland Low system, no deep-water may have formed north of the Faroes, and sea surface temperatures in the Greenland-Norwegian basin would have been colder with a more stable atmosphere than today. A large fraction of sub polar storm paths therefore would have been located slightly to the south of Iceland where a stronger oceanic temperature gradient would have been guiding the jet stream. Storm paths could then have extended across northern Europe and southern Scandinavia between latitudes 55° N and 60° N, and such storms would have carried a supply of water vapor into the central part of the Eurasian ice sheet, ensuring its growth to the maximum extent of the Drenthe substage glaciation.

5.4 Isolation of the ice-free polar ocean

Unlike the Mediterranean Sea, the polar ocean had a buffer in the form of the Greenland-Norwegian Sea basin that lies between the polar and the world oceans. As the Drenthe glaciation built up to its extreme at 142 ka and the dense fractionated water accumulated in the polar ocean, the buffer basin would have shared the accumulation, and minimized the release of the dense water to the world ocean. Unlike the Strait of Gibraltar with its depth of 285 m, no sharp density gradients would likely have been present within the Fram Strait (Fig. 5) with its maximum depth of 3700 m. Densities on both sides of the strait would have been similar, and a rapid exchange of water masses would not have occurred with the counterclockwise polar circulation. With open water over the deep polar ocean and no insulating pack ice, a lower level polar atmosphere that was warmer in winter than today is implied. As argued in Sect. 5.2, the temperature contrast between the warmer atmosphere over the open ocean and the adjacent colder lands would have generated frequent eastward-moving storm systems with heavy precipitation over adjacent high Arctic lands. Evidence for such storms is found on the northern tip of Greenland in Peary Land and on Ellesmere Island to the west. On Ellesmere and on Peary Land (Fig. 6) much of the land is barren and ice-free because precipitation rates today are quite low. But the coasts are deeply indented by old fiords formed by ice streams flowing into the sea. U-shaped valleys are abundant, indicating that thick ice sheets had covered the land and were fed by precipitation from storms passing by on the open polar ocean.

Figure 6 Google Earth image of Peary Land on the northern tip of Greenland. Image width is about 250 km.

The maximum storm turbulence would have occurred in zones not far from the coasts where the largest temperature contrasts would have been found. There evaporation and fractionation rates would have been high causing the sea surface salinity to increase. The resulting denser and saltier water would tend to sink into the deepest areas and be replaced by water upwelling from outside the zone. The sinking water would tend to slowly fill the polar basin with saltier water enriched in $H_2{}^{18}O$. The deepest water enriched in $H_2{}^{18}O$ would also accumulate in the Greenland-Norwegian basin by fractionation and by slow deep drift from the polar ocean southward through the Fram strait. The salinities and concentrations of $H_2{}^{18}O$ in the two basins at levels below the present ~480 m-depth of the Iceland-Faroe ridge might not have been greatly different. The net loss by evaporation in the polar ocean would have been replaced by a slow drift of less dense water from the Greenland basin through the Fram Strait at higher levels. Eventually a state of equilibrium circulation would prevail. The depths of the polar ocean and the Greenland-Norwegian basin below the Iceland-Faroe ridge would continue to fill with denser and saltier water enriched in $H_2{}^{18}O$. Evaporation losses from both basins would be replaced by slow net flow over the Faroe ridge. The enrichment of polar water in $H_2{}^{18}O$ would have been mirrored by depletion in the world oceans, and at the Drenthe glacial extreme at 142 ka the measured world ocean $\delta^{18}O$ could not have reflected the true low sea level that is suggested by the red dashed line at **B** in Figure 1.

6 Summary of proposed sea level corrections to the Saalian oceanic $\delta^{18}O$ record in Figure 1

6.1 A-B, 156 ka-142 ka

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Beginning at **A**, the rate of fractionation and sequestration of $H_2{}^{18}O$ in the polar ocean slightly exceeded the rate in the world ocean due to the continuing accumulation of Drenthe glacial ice. The sea level fall ended at an extrapolated level of -140 m and an age of 142 ka.

6.2 B-C, 142 ka-136.5 ka

The giant lake overflow by way of the Mediterranean Sea capped the northeastern North Atlantic and stopped Northern THC. This initiated world deglaciation by starving the ice sheets of moisture. The loss of all Northern THC also caused the loss of Antarctic thermohaline circulation because, without the THC and warmer Northern intermediate-level water input, the shelf ice and heavy sea ice expands out over deep water where convection prevents the seasonal freezing that otherwise results in deep water formation (Johnson, 2015b). Therefore with the loss of the two main sources of deep water formation the world ocean became more stratified, and instead of the modern world ocean mixing time of about a thousand years, the mixing time increased to many thousands of years. The depressed $\delta^{18}O$ at 136.5 ka in Figure 1 might be explained if a modest THC was restored quite late in the deglaciation before much world mixing had occurred. Most of the melt water would still have been found in the upper levels of the ocean. The formerly sequestered $H_2{}^{18}O$ would have been even more concentrated near the surface where it would have had little or no effect on the benthic record at that time. The mixing delay could therefore explain the small rise in $\delta^{18}O$ in the V28-238 record from 142 ka to 136.5 ka. The larger concentration of $H_2{}^{18}O$ near the ocean surface and the cold temperatures associated with the incomplete Northern deglaciation would explain the even greater suppression of $\delta^{18}O$ found in typical planktonic sediment records, making the Drenthe deglaciation almost invisible in both surface and benthic $\delta^{18}O$ records.

6.3 C-D, 136.5 ka-130 ka

Most of the following Warthe glaciation occurred under renewed fractionation in the polar ocean because the Siberian rivers again became ice-blocked. The Warthe sea level fell to about -80 m as implied by the shallow-water corals that had grown in Aladdin's Cave on presently uplifted New Guinea (Esat et al., 1999). Therefore, somewhat as in the **A-B** interval, the sequestration of $H_2{}^{18}O$ during the shorter Warthe interval raised the inferred sea level at **G** to about 30 m above the likely true level.

6.4 D-E, 130 ka-126 ka

The end point sea level of this interval at **E** in Figure 1 at the beginning of the last interglacial was several meters above present (Bender et al.,1979; Chen et al., 1991), which is roughly consistent with the total $\delta^{18}O$ change. However, the details of the physical changes in the interval **D-E** like those in **B-C** are less clear. The Warthe deglaciation began when strong African monsoons shut down the THC with low salinity Mediterranean water, but the interval ended with Northern glacial ice gone and all the sequestered $H_2{}^{18}O$ well mixed into the world ocean, implying strong THC. There is a precedent for such a switch to strong THC early in the last deglaciation. Bond et al. (1993) reported an abrupt positive jump of almost 5 ‰ in the $\delta^{18}O$ of the Greenland GRIP ice core at 14,700 calendar years before present (Fig. 7). Sea level then, as measured at the Barbados site, was still low at about -100 m (Figure 2 in Fairbanks, 1989). At about that time an abrupt sea level rise of 30 m began that could only have been caused by a massive loss of Antarctic ice. The δ^{18}O change in the GRIP ice core was probably caused by the restoration of strong North Atlantic THC.

However, with the linkage of the deep ocean oscillation (Johnson, 2015b), it is difficult to say if a collapse of Antarctic ice mass was the cause or the effect of the restored northern THC, although the former seems more likely. A collapse of Antarctic marine-based ice would have to have occurred early in the Warthe deglaciation to have produced the complete mixing result, whereas in the Drenthe deglaciation a similar collapse must have occurred late in the deglaciation before world ocean mixing became very significant.

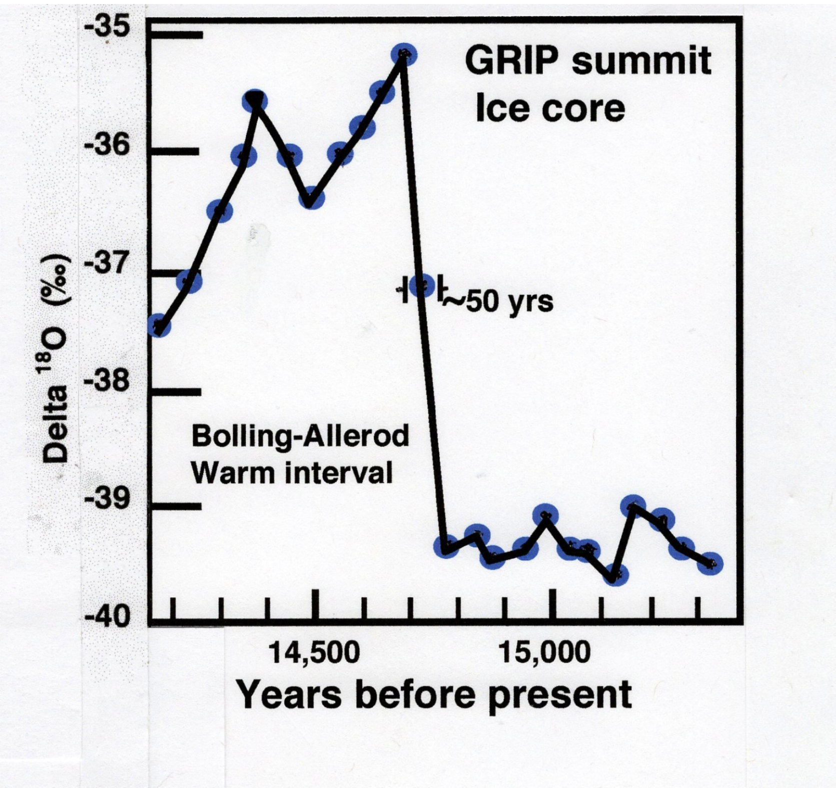

Figure 7 **A large step change in the δ^{18}O record of a Greenland ice core, interpreted as the result of the sudden restoration of strong thermohaline circulation in the Greenland-Norwegian Sea early in the last deglaciation. Modified from Bond et al. (1993).**

7 Discussion and conclusions

7.1 The rapidity of the Warthe reglaciation: The ice-free Baffin Bay model

The sea level fall from +4 m to almost -80 m between 136.5 ka and 130 ka was remarkably rapid. A factor that favored the rapid fall was the residual ice in Eurasia and in Canada that remained when the giant Siberian lake was drained away and low-salinity oceanic capping ceased. But to explain the sea level fall of about 80 m in 6,500 yrs, a mechanism is needed that favors glacial accumulation on both Northern continents and on the Antarctic continent. The ice-free Baffin Bay scheme proposed for rapid ice accumulation at the initiation of the last ice age at 120 ka (Johnson, 2015a) would accomplish this if slightly warmer deep-water formation could be maintained in Baffin Bay.

In the conceptual model for initiation of the last ice age as argued in Sect. 5, a persistent storm system developed over the warmer and perennially open water in the Labrador Sea and Baffin Bay, caused by loss of stratification in Baffin Bay at 120 ka. Strong evidence has been reported (Koerner et al., 1988) for open water at the north end of Baffin Bay when the last ice age began. They found willow pollen at the base of a core from Devon Island ice cap in a region where no willows grow today. This observation is consistent with the report by Fillon (1985) of warmer water foraminifera in deep-sea sediment at the start of the last ice age in core HU75-58 located 200 km east of the southern tip of Baffin Island where only polar water species are found today. These observations imply a persistent cyclonic storm system centered over the Labrador Sea, which brought an order of magnitude increase in snowfall to northern Canada and Greenland (Johnson, 2015a). The strength of this cyclonic system also probably prevented the development of the adjacent Iceland Low pressure system, which consequently cooled Scandinavia and northern Europe and caused a southward shift of the stronger oceanic temperature gradients. Storms were then steered over the frozen Barents Sea and the Scandinavian ice sheet nucleation areas. The open water in the Labrador Sea west of Greenland was maintained for ~500 yrs, during which world sea level fell by 2.4 m as measured on uplifted Barbados (Johnson, 2001; Johnson 2015a; Johnson 2015b). The open water in the Labrador Sea and Baffin Bay was maintained by a strong and saline West Greenland Current flow that enabled deep water formation to the north in Baffin Bay despite fresh water input to the Labrador Sea from the Hudson Strait outflow. The low salinity of that outflow was temporarily counteracted by a more saline Irminger Current input into the West Greenland Current that was weakened or lost after ~500 yrs because the deep ocean oscillation (Johnson, 2015b) reduced the Northern THC.

In the context of the Warthe glaciation, the deep water formed in the northern part of ice-free Baffin Bay would have reached the Southern Ocean and limited the formation of sea ice and shelf ice, thus allowing the circumpolar storm paths to remain close to the coast to increase the accumulation rate of glacial ice on the Antarctic continent. However, to make this concept work continuously for the ~6,000 yr-long Warthe glaciation, during which the Irminger current would often have been weak, the Hudson Strait fresh water outflow that lowers Labrador Sea salinity must have been blocked from the start of the new Canadian glaciation to ensure the higher salinity needed for deep-water formation in Baffin Bay. There is a recent analog for the blockage of Hudson Strait late in the last deglaciation when the North American ice sheet had become much smaller. At 10.9 ka (calendar years) a re-advance of the Laurentide ice sheet across Ungava Bay extended as far to the north as the Hall Peninsula and completely blocked Hudson Strait (Stravers et al, 1992). Blockage of Hudson Strait after several thousand years of early ice sheet growth is also suggested by the model of Andrews and Mahaffy (1976). A similar lasting blockage of Hudson Strait may have facilitated an ice-free Labrador Sea and the Warthe glaciation. Blockage may have also triggered the renewed glaciation at about 200 ka and the subsequent ~55 m fall of sea level that ended about 193 ka (Fig. 1). Therefore an ice-

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blocked Hudson Strait and an ice-free Labrador Sea and Baffin Bay may have occasionally initiated long intervals of uninterrupted ice sheet accumulation and rapid world sea level fall during the Pleistocene.

7.2 Conclusions

The consequences of fractionation and sequestration of $H_2{}^{18}O$ in the ice-free polar ocean remove much of the confusion associated with evidence for events that occurred in the world climate records during Termination 2. The root cause of the fractionation was the extreme extent of Eurasian glacial ice that blocked Siberian river flow into the polar ocean, and this is consistent with the concept that the moisture supply is the dominant factor in the growth or shrinkage of glacial ice sheets. During the Pleistocene, the Mediterranean salinity and strong African monsoons that are controlled by orbital precession may play a decisive role in reducing the moisture supply and ensuring the nearly complete removal of Northern Hemisphere glacial ice.

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