The Deep Ocean During the Last Interglacial Period

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Oxygen isotope analysis of benthic foraminifera in deep sea cores from the Atlantic and Southern Oceans shows that during the last interglacial period, North Atlantic Deep Water (NADW) was 0.4° ± 0.2°C warmer than today, whereas Antarctic Bottom Water temperatures were unchanged. Model simulations show that this distribution of deep water temperatures can be explained as a response of the ocean to forcing by high-latitude insolation. The warming of NADW was transferred to the Circumpolar Deep Water, providing additional heat around Antarctica, which may have been responsible for partial melting of the West Antarctic Ice Sheet.

The climate of the last interglacial (LIG) period, from 129,000 to 118,000 years ago (*1*, *2*), was slightly warmer than today's and is often viewed as an analog of the climate expected during the next few centuries. Recent assessments of the LIG climate have provided strong evidence that sea level was 4 to 6 m above the present level, due to partial melting of both Greenland and the West Antarctic Ice Sheet (WAIS) (*3*, *4*). At peak interglacial conditions, summer temperatures were 2° to 5°C warmer than today in the North Atlantic (*5*) over Greenland (*6*) and the Arctic (*7*). The Norwegian-Greenland Sea experienced large variability, but during the warmest period, the Arctic oceanic front was located west of its present location (*8*). Consequently, the Arctic climate was warm enough to explain the shrinking of the Greenland Ice Sheet during the LIG (*9*).

In the Southern Hemisphere, an ~2°C warming occurred over the Antarctic Plateau during the LIG (*10*), but it could not have resulted in any melting because local air temperature was still extremely cold (~–50°C). In the Southern Ocean, summer sea surface temperatures were about 2°C higher than during the Holocene (*11*, *12*). Over New Zealand and Tasmania, the LIG warming was between 0° and 2°C (*13*, *14*). Such increases in surface water or air temperature seem too small to have resulted in substantial melting of the WAIS (*15*).

However, the WAIS is sensitive to deep ocean temperatures (*16*). Indeed, the volume of this ice sheet is related to the efficiency of ice shelves in blocking the ice flow from the central part of the ice sheet outward, and ice shelves are themselves sensitive to deep ocean temperatures. Under modern conditions, the meridional overturning circulation carries warm surface waters northward in the Atlantic, replacing the export of the relatively cold North Atlantic Deep Water (NADW). However, NADW is warmer than the cold bottom water formed around Antarctica. NADW therefore carries heat to the Southern Ocean. It mixes with recirculated deep water from the Indian and Pacific Oceans, forming a relatively warm deep water mass, the Circumpolar Deep Water (CDW), characterized by a temperature maximum of ~+2°C around 600 m depth and overlying the cold Antarctic Bottom Water. CDW floods the floor of the Amundsen and Bellinghausen Sea continental shelves and reaches the undersides of ice shelves flowing from the WAIS. This heat flow, which originates from the Northern Hemisphere, results in ice-shelf bottom melting near grounding lines (*17*). This melting process substantially contributes to decreasing the stability of the ice shelves that drain the grounded part of the WAIS (*18*).

To investigate whether the LIG ocean could have helped to destabilize the West Antarctic ice shelves and ice sheet, we examined the oxygen isotopic composition ($\delta^{18}O$) of benthic foraminifera. This quantity is a function of both temperature and seawater $\delta^{18}O$. On the one hand, the isotopic composition directly reflects isotopic variations in the ambient water (δw). On the other, the fractionation between the carbonate shell and the water increases by about 0.25 per mil (‰) for each degree that the water is cooled. Because the global ocean circulation during the LIG period was very similar to the present circulation (*19*), $\delta^{18}O$ differences between benthic foraminifera from the LIG period and from the Holocene in a core reflect both the temperature and $\delta^{18}O$ differences experienced by the same deep water mass as the one that is present today at the core location.

The LIG $\delta^{18}O$ records are characterized by quasi-constant values forming a plateau, in agreement with coral reef data showing that the LIG high–sea-level period lasted at least 7000 years (*1*, *2*, *20*). In all cores, the LIG plateau exhibits $\delta^{18}O$ values that are significantly lighter than those of the Holocene (Fig. 1). To accurately determine the $\delta^{18}O$ difference between these two

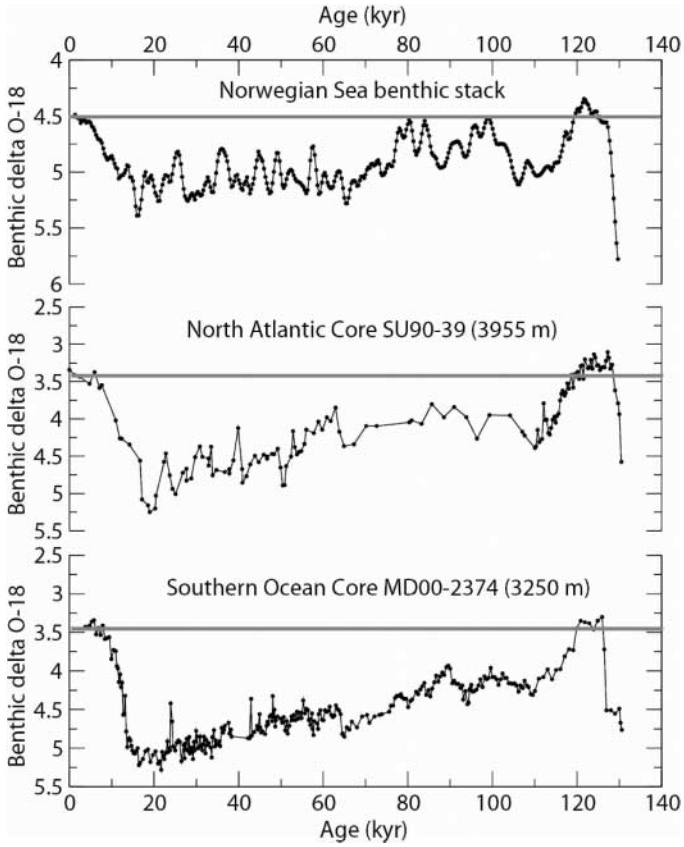

Fig. 1. Comparison of three $\delta^{18}O$ benthic foraminifer records for the past 135,000 years, obtained from deep-sea cores raised from the major water masses found in the Atlantic and Southern Oceans. The upper panel shows a Norwegian Sea benthic stack record (*36*), the middle panel is core SU 90-39 (52°32′N, 21°56′W, depth 3955 m) from the North Atlantic, and the lower panel is core MD 00-2374 (46°04′S, 96°48′E, depth 3250 m) from the Southern Ocean. All records encompass the LIG (135,000 to 110,000 years ago) and the present interglaciation (since 8000 years ago), together with the last glacial period. The gray line in each panel shows the mean Holocene $\delta^{18}O$ value. kyr, thousand years.

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interglacial periods, we compared the records of benthic foraminifera from the same species in 42 cores (table S1). The mean $\delta^{18}O$ value of LIG benthic foraminifera in these cores is 0.11 ± 0.01‰ lighter than during the Holocene, but with significant regional variations: The $\delta^{18}O$ difference between the LIG period and the Holocene varies from 0.15 ± 0.02‰ in the Norwegian Sea to 0.11 ± 0.01‰ in the Atlantic Ocean and to only 0.04 ± 0.03‰ in the Southern Ocean (table S2).

Light $\delta^{18}O$ values of LIG benthic foraminifera are partly explained by partial melting of both Greenland and the WAIS, which injected isotopically light meltwater into the ocean. To estimate the seawater $\delta^{18}O$ variation associated with LIG sea level, which was 4 to 6 m above the present level, we note that between the last glacial maximum and today, the sea level rose by 130 ± 10 m (*21*) and seawater $\delta^{18}O$ decreased by 1.05 ± 0.20‰ (*22*, *23*). This implies that seawater $\delta^{18}O$ decreases by 0.008 ± 0.002‰ per meter of global sea-level rise. This rate is likely to be an upper estimate for interglacial conditions, because ice melting at the end of the glaciation originated from snow deposited at low temperature and highly depleted in ^{18}O. In contrast, ice melting during an interglacial climate would originate from snow deposited at temperatures higher than those of glacial climate and less depleted in ^{18}O. We conclude that the 4- to 6-m LIG sea-level rise would have resulted in a seawater $\delta^{18}O$ decrease of 0.03 to 0.06‰. This change was uniform in all water masses, because the turnover of the ocean (about 1500 years) is much shorter than the LIG duration. Therefore it should be recorded in all benthic foraminiferal $\delta^{18}O$ values. We then computed a local $\delta^{18}O$ anomaly as the difference between the measured foraminiferal $\delta^{18}O$ value and that owing to the input of meltwater. This local $\delta^{18}O$ anomaly is only due to changes in deep water $\delta^{18}O$ and temperature. For the deep Southern Ocean (1900 to 3500 m), the anomaly is not statistically different from zero. This result implies that during the LIG, bottom water formed on the Antarctic continental shelf at the freezing point, as it does today. In the Norwegian Sea and the Atlantic Ocean, the anomaly is significantly different from zero at the 1σ level (–0.15 and –0.11‰, respectively). However, it is small and can be considered to be a minor departure from modern conditions. Such small changes in the physical properties of water masses tend to develop at almost constant density, as observed in hydrographic data

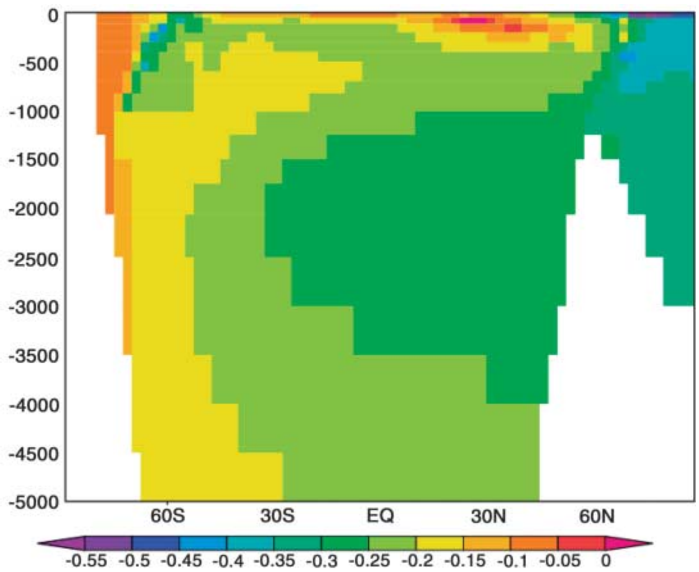

Fig. 2. Calcite $\delta^{18}O$ anomaly (LIG – modern) simulated by CLIMBER-2 (in per mil) in the Atlantic Ocean (computed between the averages over the last 100 years of the simulations). Calcite $\delta^{18}O$ is computed as $\delta^{18}O_c = 21.9 - 0.27 + \delta^{18}O_w - \sqrt{310.61 + 10T}$ where *T* is the oceanic temperature (*37*). The global $\delta^{18}O$ change in seawater composition of –0.045‰ due to sea-level change at the LIG is taken into account.

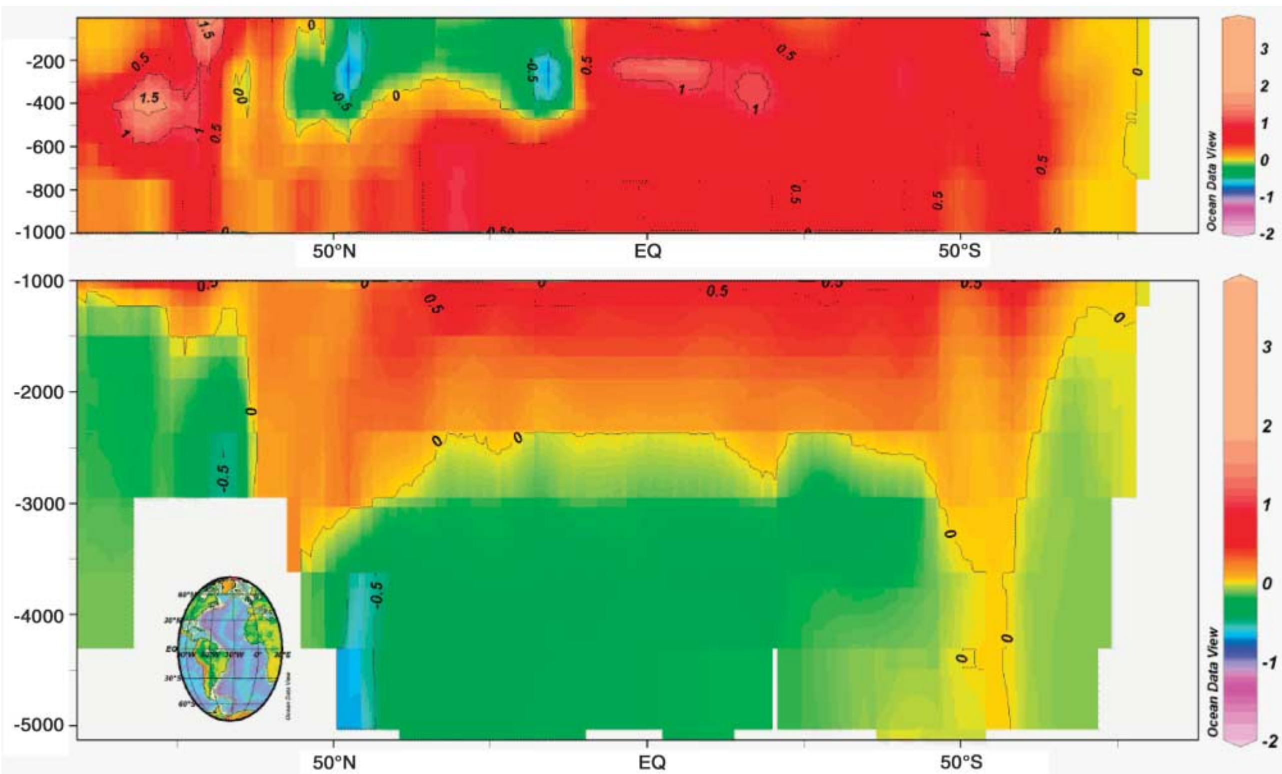

Fig. 3. Annual mean temperature anomalies (LIG – modern) simulated by LOVECLIM (in °C) along the western boundary of the Atlantic Ocean. The field is averaged over the last 100 years of the simulation. The inset map shows the path chosen for the section.

collected during the past 50 years (*24*, *25*). At present, the temperature and salinity of NADW are close to 3°C and 34.96 practical salinity units (psu). To maintain a constant density, salinity must increase by 0.12 psu for each degree that the water is warmed. Thus, we can compute that the $\delta^{18}O$ anomalies measured in LIG benthic foraminifera result from a 0.37° ± 0.20°C warming of NADW, compensated for by a 0.04 increase in salinity. Similarly, a 0.57° ± 0.2°C warming compensated for by a salinity increase of 0.07 psu occurred in the Norwegian Sea (table S2).

Among the factors responsible for a LIG warmer than today, insolation changes are the best candidates. Indeed, Earth's orbital parameters during the LIG favored warm Northern Hemisphere summers as compared to the present. LIG atmospheric greenhouse gas concentrations were not significantly different from preindustrial values. To simulate the LIG atmosphere/ocean state, we therefore forced our climate models with the insolation values for 126,000 years before the present (supporting online text). We used two Earth System models of intermediate complexity: CLIMBER-2 (*26*) and LOVECLIM (*27*). In CLIMBER-2, the ocean is represented by three latitude/depth basins, and the computed calcite $\delta^{18}O$ can be directly compared with measurements. LOVECLIM includes a full-ocean general circulation model but no oceanic oxygen isotope calculation.

The CLIMBER-2 model simulates negative calcite $\delta^{18}O$ anomalies (LIG –control) in all ocean basins (Fig. 2). The magnitude of these anomalies decreases from the Nordic Seas (–0.16‰), through the Atlantic (–0.12‰), into the Southern Ocean (–0.06‰). There is good agreement with data (table S2), in particular for the Atlantic Ocean, where the reconstructions are the most robust. Furthermore, both models simulate sea surface temperature changes consistent with reconstructions and numerical studies for the LIG (*9*, *28–30*) (supporting online text). Sea surface temperatures increase at nearly all latitudes, with a maximum increase at high northern latitudes and a secondary maximum in the Southern Ocean. Maxima occur at high latitudes because of reduced sea-ice cover. The increased summer boreal insolation is responsible for a significant melting of the northern sea ice, which translates into a year-round warming of the ocean due to its large thermal inertia. As a consequence of these higher sea surface temperatures, ocean evaporation increases. This increase is not compensated for by an increase in precipitation. Consequently, salinity increases, which agrees with the paleoceanographic reconstructions (*5*). Hence, the model results are in broad agreement with the surface hydrographical reconstructions and the benthic $\delta^{18}O$ data, the latter being simply interpreted as the response of the ocean to the LIG insolation values.

Simulated density is quasi-constant because the impacts of the temperature and salinity changes compensate for each other (table S2). Both models therefore confirm the hypothesis at the basis of our interpretation of the benthic $\delta^{18}O$ data. Furthermore, a nearly constant density explains why the deep oceanic circulation is little affected by the insolation changes at 126,000 years before the present. Nordic Sea surface waters, fed by warmer and saltier North Atlantic surface waters, form a deep water mass, also slightly warmer and saltier than today [change in temperature (ΔT) = 0.6°C and change in salinity (ΔS) = 0.04 psu for the CLIMBER-2 model; ΔT = 0.5°C and ΔS = 0.06 psu for the LOVECLIM model, see table S2]. This water invades the North Atlantic as a large NADW mass (Fig. 3), which then flows to the Southern Ocean.

As a result, the Antarctic Circumpolar Waters are slightly warmer than today. Our LIG simulations show a ~0.1° to 0.5°C warming in the upper 500 m of the Southern Ocean close to the Antarctic coast (Fig. 3 and fig. S2). This value is similar to the 0.5°C warming simulated by the National Center for Atmospheric Research model for the upper 200 m of the Southern Ocean (*3*). Although apparently modest, this warming should not be underestimated. Recent results obtained with satellite radar interferometry reveal that bottom melting near an ice-shelf grounding line is strongly influenced by the temperature of the surrounding seawater, and that the melting rate at the base of ice shelves increases by 1 m per year for each 0.1°C rise in ocean temperature (*16*). Thus, in addition to the higher sea level resulting from the partial melting of the Greenland ice sheet, the 0.1° to 0.5°C warming that we have estimated for LIG NADW and CDW may have affected vulnerable WAIS grounding lines and further weakened the ice shelves by causing thinning from below.

Our data show that changes in climate in the high-latitude North Atlantic could have triggered some ice sheet melting in Antarctica, but they provide no information on the speed at which the WAIS shrank during the LIG. Although it is not our goal to predict the future of the WAIS, we note that recent ocean temperatures directly seaward of Antarctica's continental shelf have already increased by ~0.2°C (*31*), a warming comparable to that of the LIG period. Consequently, the future evolution of the WAIS might be a key component of sea-level change resulting from anthropogenic warming, as Mercer warned more than 25 years ago (*32*).

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Supporting Online Material

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SOM Text
Figs. S1 and S2
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References

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