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Examining deglacial circulation in the North Atlantic

SGHK8

1 Exercise 1 - Mid-depth temperature and $\delta^{18}O_{SW}$ reconstruction

1.1 Question 1

According to Barker *et al.* [1], foraminiferal Fe/Mg molar ratios greater than approximately 0.1 are indicative of potential contamination by Mg-rich clay silicates. Mg/Ca measurements could therefore be rejected if they do not meet the criteria:

$$\frac{\mathrm{Fe/Ca}}{\mathrm{Mg/Ca}} < 0.1\,\mathrm{mol}\cdot\mathrm{mol}^{-1} \tag{1}$$

However, visual assessment of the magnitude of potential contamination (Figure 1) shows that all but one of the eight measurements failing the this criteria are close to the rest of the data in terms of Fe/Mg. As a result, only the 16.39 ka measurement labelled in Figure 1 is rejected for contamination.

1.2 Question 2

As shown in Figure 1.2, the mid-depth seawater oxygen isotope excursion $\delta^{18}O_{SW}$ declines to negative values after the Last Glacial Maximum (LGM) in correspondance with falling water temperatures. The two trends fall sharply toward the beginning of Heinrich Stadial 1 (HS1) before diverging: in contrast to the gradual temperature recovery from nearly 0 °C, $\delta^{18}O_{SW}$ continues to strongly decrease until the transition into the Bølling–Allerød warm period (BA, 14.7–12.9 ka [2]), at which point both lines rise together rapidly. The Younger Dryas (YD) marks a very abrupt return to HS1 temperatures, although the concurrent drop in $\delta^{18}O_{SW}$ is not as severe, whereas each record shows relative stability over the course of the Holocene.

The imprint of local salinity on foramineferal Mg/Ca may result in significant systematic positive bias in reconstructed $\delta^{18}O_{SW}$ [3], whereas uncertainties in foramineferal $\delta^{18}O_{\mathrm{carbonate}}$ stemming from nonclimatic factors, such as measurement imprecision or variations in seawater carbonate ion concentration, are comparatively small [4]. Assuming these issues are negligeable, the control of water oxygen isotopic composition hydrological fractionation processes at the surface make it useful proxy for salinity, from which an interior ocean water mass's surface origin can be inferred [5, 6]. Although sea ice formation may be an important control on intermediate water salinity, given the subpolar location

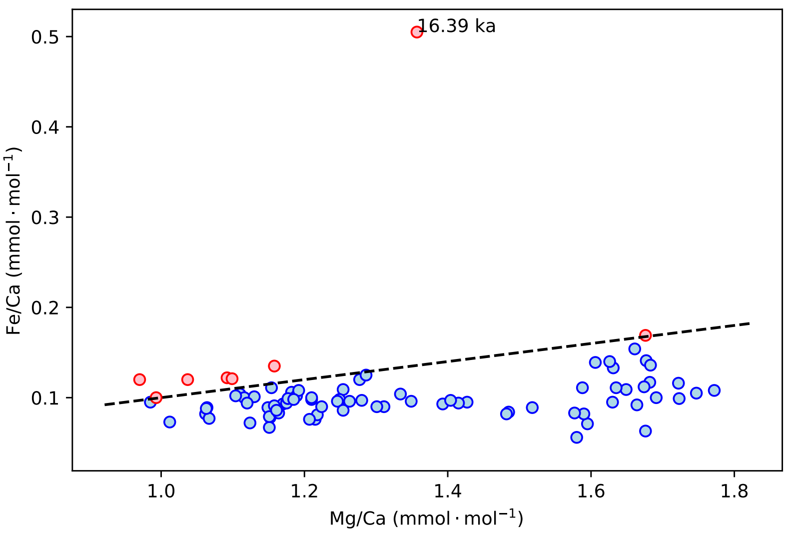

Figure 1: Ratio of Mg/Ca to Fe/Ca in benthic foraminifera *M. barleeanuum* in intermediate sediment core RAPiD-10-1P, on the South Iceland Rise. Dashed line seperates potentially contaminated measurements (red points) according to Equation 1. Labelled point is identified for rejection.

of the RAPiD-10-1P core, the relatively small isotopic effect of brine rejection means that the surface source $\delta^{18}O_{SW}$ signal is largely unaltered [7]. Therefore, in accordance with similiar studies [e.g. 2, 8], the changes in $\delta^{18}O_{SW}$ and Temperature over HS1 (Figure 1.2) could be interpreted to show the displacement of cold, hypersaline, $\delta^{18}O$-enriched polar waters, originating from open-ocean convection, by an increasing sink of warmer, isotopically lighter glacial meltwater due to enhancement of local brine formation, while the abrupt drops preceeding HS1 appear to record large-scale freshening events. These results support previous studies hypothesising the suppression of traditional North Atlantic Deep Water (NADW) during deglacial stadials by intense ocean stratification associated with the massive input of freshwater into the North Atlantic from the Northern Hemispheric ice sheets [9, 10].

2 Exercise 2: Tracers of deep ocean circulation change

2.1 Question 1

See Figure 3

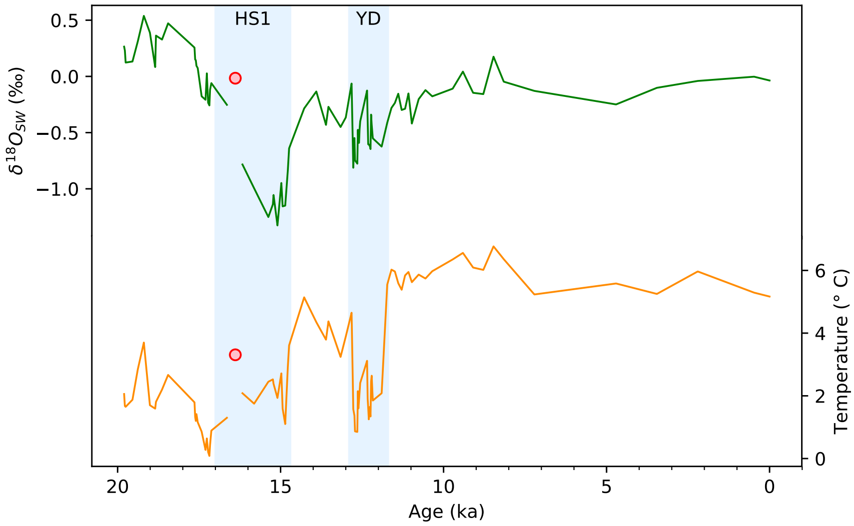

Figure 2: Trends in the ice-volume corrected $\delta^{18}\mathrm{O}_{\mathrm{SW}}$ (top) and mid-depth water temperature since the Last Glacial Maximum determined from Mg/Ca palaeothermetry (bottom) recorded in RAPid-10-1P. Red points and the gap in the timeseries signify a rejected measurement (see text in Section 1.1). Shaded regions mark Heinrich Stadial 1 (HS1) and Younger Dryas (YD), as defined in Thornalley *et al.* [2].

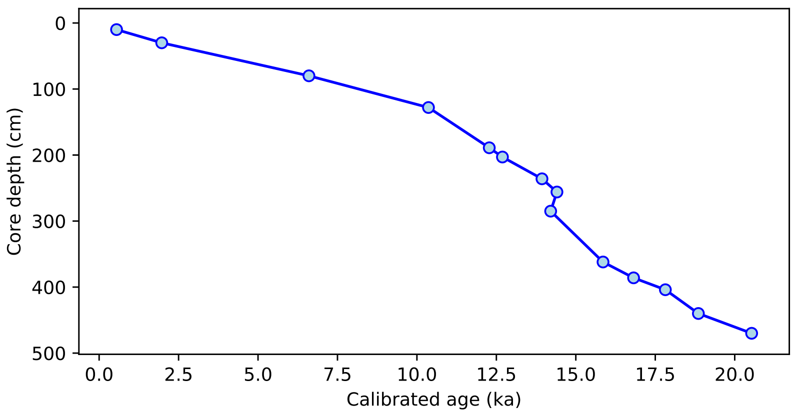

Figure 3: Planktic foraminifera calibrated age against depth in sediment core OCE-326-GGC14, on the Laurentian fan.

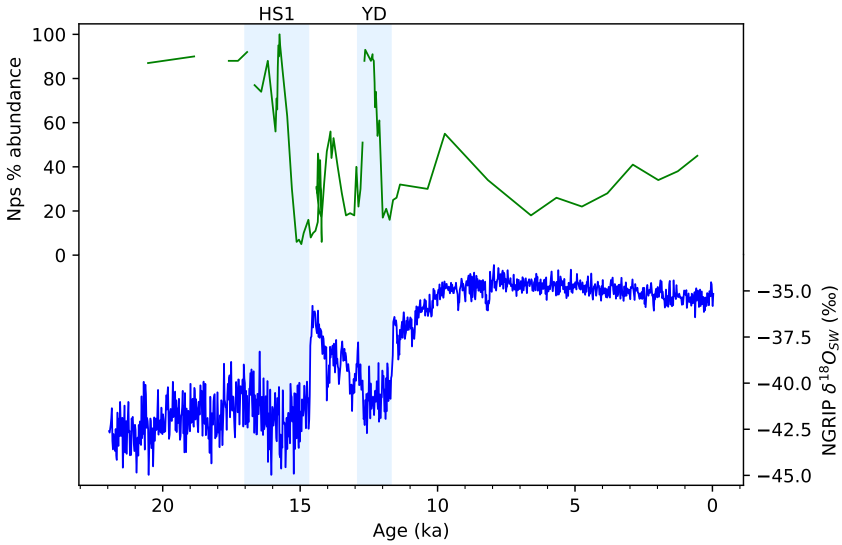

Figure 4: Time series of abundance of planktic foraminifera *N. pachyderma* sinistral (*Nps*) in OCE-326-GGC14 (top) and oxygen isotope excursion in Greenland ice core NGRIP (bottom).

2.2 Question 2

Figure 4 shows a strong antiphase relationship between *Nps* abundance in the North Atlantic and δ^{18}O in Greenland, and both record abrupt climate cooling during the HS1 and YD stadials. While changes in the latter appear to lag those in the former by approximately 0.5–1 ka, there is uncertainty in the calibration of foraminiferal ^{14}C radiocarbon dates due to measurement error and, in particular, the assumption of a constant 400 year radiocarbon reservoir age — a value that was likely higher during stadial events due to decreased ocean ventilation [11, 12].

2.3 Question 3

The benthic–planktic foraminifera ^{14}C offset (Figure 5, top) is considerably larger at LGM, HS1 and YD than during the BA and the Holocene. Since $\Delta^{14}\mathrm{C}_{\mathrm{B-P}}$ of a water mass effectively records the ventilation age (i.e. time since last exposure at the surface) [6], these results indicate that the North Atlantic was highly stratified during the cold periods, with younger northern sourced waters displaced in the deep ocean by the intrusion of more ancient waters of southern origin. This supports evidence for the slowdown or shutoff of NADW formation during these cold periods, allowing Antarctic Bottom Water to fill the global abyssal basins [11, 15, 16]. Likewise, the rapid rejuvenation of deep waters at the HS1–BA and YD–Holocene transitions shown in the $\Delta^{14}\mathrm{C}_{\mathrm{B-P}}$ time series indicates rapid resumption of the contemporary (warm) mode ocean overturning circulation.

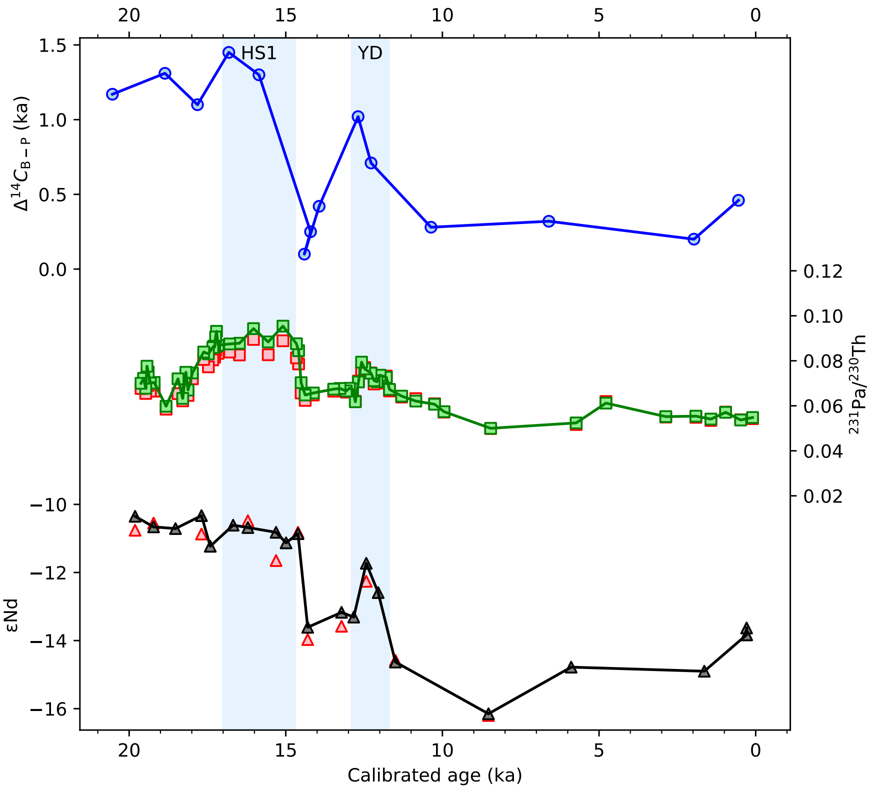

Figure 5: Time series of benthic–planktic foraminifera ^{14}C ventilation age difference in OCE-326-GGC14 (top); $^{231}Pa/^{230}Th$ ratio in calculated against ^{238}U (green) and ^{232}U (red) OCE326-GGC5 core on Bermuda rise [data from 13]; Neodynium isotopic ratio excursion in unclean foraminifera (black) and reductively cleaned fish debris (red) from OCE326-GGC6 sediment core on Bermuda Rise (bottom) [data from 14].

2.4 Question 4

The $\Delta^{14}C_{B-P}$ changes described in Section 2.3 are largely consistent with isotopic trends described in other studies, as depicted in Figure 5. The ratio of ^{231}P sequestered in marine sediments to the more quickly scavenged ^{230}Th is an indicator of the residence time of the overlying water mass [6], and the high values recorded on the Bermuda rise for the HS1, and to a lesser extent YD, indicate poor export of ^{231}P from the region during the postglacial stadials. This can be explained by the arrest of the southward movement of NADW, as described above, with the steep $^{231}Pa/^{230}Th$ drop at ~15.5 marking the abrupt resumption of the AMOC following HS1 [13]. However, one issue with this using the particular proxy to infer changes in global circulation is that the distinction between a complete shutdown and a more moderate shift to a shallower, more locally confined circulation of North Atlantic waters is not always readily apparent, since the sedimentary $^{231}Pa/^{230}Th$ signal essentially records change in the entire column of water above [14].

Since $^{143}Nd/^{144}Nd$ (ϵNd) serves as a form of isotopic signature of different ocean mass sources, it can be used to supplement the interpretation of the other records in Figure 5 [6]. The high values from the LGM through to the end of HS1 provide evidence for the incursion of AABW into the Northwest Atlantic, with a marked shift to less radiogenic Northern-sourced waters following the YD, supporting the theory of AMOC modulation as a distinctive feature of deglacial climate change [14].

2.5 Question 5

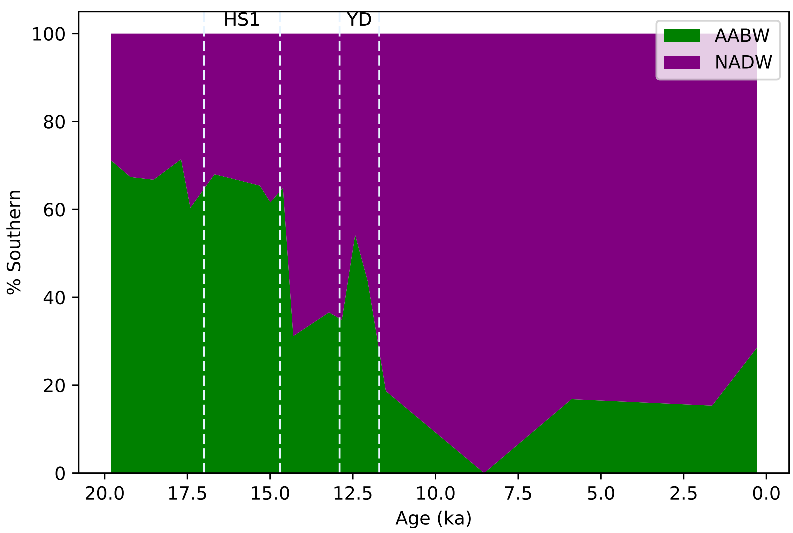

Figure 6: Proportion of northern (NADW) and southern (AABW) waters at Bermuda Rise , as estimated from unclean foraminifera ϵNd in OCE326-GGC6 sediment core.

To visualise the varying influence of Northern and Southern sourced waters over the past 20 ka, an assumption is made that AABW (ϵNd$=-8$ [17]) and NADW (ϵNd$=-16.15$, lowest value in core) represent the two mixing end-members for the ocean mass bathing the Bermuda rise, as recorded in sediment core OCE326-GGC6. The dominance of NADW during warm periods and the intrusion of southern sourced waters during stadials can clearly be seen in Figure 6. However, the exact values used here were chosen rather abitrarily from among a large range of observations for the *present*, whereas values may have varied considerably over time, especially given that AABW, treated here as an opposing end-member to NADW, is in fact a product of Atlantic, Indian and Pacific waters which are likely to have mixed in different ratios in the past. Futhermore, the dominant processes controlling water radiogenic properties may have changed. This highlights the uncertainty introduced when certain ocean mass properties, such as ϵNd, are assumed to be constant over time.

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