Sea-level reversal during Termination II

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ABSTRACT

The termination of the penultimate glacial period (TII) shows both similarities and differences to the last termination (TI). Both terminations show significant cold reversals in the postglacial warming trend. TI consists of a continuously increasing sea-level trend, whereas TII may demonstrate a sea-level reduction midway through the termination. We present a new, continuous sea-level record for TII derived from Red Sea δ^{18}O records that supports the existence of the TII sea-level reversal. The record gives an unprecedented look at the structure of the TII sea-level reversal, which consists of an early highstand lasting several millennia in duration, followed by a 30 ± 12 m to 40 ± 12 m reduction in sea level and a stillstand of several millennia, before the final sea-level rise to the marine oxygen isotope stage (MIS) 5e interglacial. We suggest that there is a link between the sea-level reversal and internal, millennial-scale variability in the climate system.

Keywords: Termination 2, MIS 6, MIS 5e, sea level, Red Sea.

BACKGROUND

During glacial times, large fluctuations in the size of the continental ice sheets occurred on time scales of thousands (Siddall et al., 2003; Cutler et al., 2003; Potter and Lambeck, 2004; Thompson and Goldstein, 2005) to hundreds of thousands of years (e.g., Chappell and Shackleton, 1986; Shackleton, 1987; Lambeck and Chappell, 2001). The termination of the last glacial period, named Termination I (TI), was characterized by several episodes of rapidly rising sea level known as meltwater pulses. The meltwater pulses contributed to a monotonic sea-level rise during TI, i.e., there were no significant intercalated episodes of sea-level fall, even though temperatures across much of the Northern Hemisphere reverted back to near-glacial conditions during the Younger Dryas period halfway through the deglaciation (e.g., Fairbanks, 1989; Bard et al., 1990; Grootes et al., 1993; Arz et al., 2003). Although no evidence for Younger Dryas sea-level fall is available, evidence does exist for the advance of the Fennoscandian ice sheet during the Younger Dryas (Mangerud et al., 1979).

Like TI, the changes in climate and ice volume linked to the termination of the penultimate glacial period (TII) were likely characterized by separate, abrupt events (Sarnthein and Tiedemann, 1990). Many Northern Hemisphere climate records suggest that TII included a Younger Dryas–like climate reversal, but there is a contradiction between different records whether this was larger or smaller in magnitude than the Younger Dryas itself (Cannariato and Kennett, 2005). Despite the apparent similarities in climate records between TI and TII, several studies present data suggesting that TII may not have consisted of a monotonic sea-level rise, but instead included an interlude of significant sea-level fall.

The first indications for a reversal in the trend of increasing sea level during TII were presented by Esat et al. (1999), who dated corals from Aladdin's Cave (Huon Peninsula) using U-Th techniques and found that the coral growth phase indicated a sea-level lowstand midway through TII. However, the Aladdin's Cave information relied on U-Th dates from a single Porites coral, and it needs further, independent corroboration to confirm a sea-level reversal. Antonioli et al. (2004) described very thin marine layers in Argentarola Cave speleothems, northwest Italy, which have been dated by U-Th (Bard et al., 2002). Marine serpulids grew on several speleothems during TII, thus probably during the early highstand. However, the marine overgrowth and subsequent speleothem layer are much too thin to provide adequate dateable material to rule out any ambiguity in this explanation or to allow dating of the TII sea-level reversal. In summary, several independent lines of evidence point toward the occurrence of a TII sea-level reversal, but each line of evidence has important weaknesses. The published evidence is either discontinuous or hard to date accurately.

Here we reconstruct a sea-level curve using new highly resolved planktonic foraminiferal δ^{18}O data through TII from central Red Sea sediment core GeoTue KL11. This reconstruction offers a strong, independent corroboration to the previous suggestions of a sea-level reversal during TII (Esat et al., 1999; Gallup et al., 2002; Antonioli et al., 2004). By portraying the sea-level reversal in an unprecedented continuous record, our results also offer new insight into the detailed structure of this sea-level reversal.

METHOD

Red Sea Sediment Records and Sea Level

The Red Sea is a semi-isolated, highly evaporative basin, which currently generates salinities within the basin of up to 40 p.s.u. (see Siddall et al., 2002, and references therein for further discussion). Such high salinity values indicate the important "strangling influence" of the shallow Hanish Sill (137 m deep; Werner and Lange, 1975; Ivanova, 1985). The sill joins the Red Sea to the Indian Ocean via the Gulf of Aden. The strangling influence of the shallow sill is especially important for periods of low sea level and strongly influences both Red Sea microfossil communities (Rohling et al., 1998; Fenton et al., 2000), and stable oxygen isotope (δ^{18}O) ratios (Hemleben et al., 1996; Siddall et al., 2003, 2004). The latter are recorded in calcareous microfossil shells, allowing researchers to recover records of sea-level change through time by analyzing microfossils from Red Sea sediment cores. The sensitivity of Red Sea δ^{18}O records to sea-level change has been modeled and the results exploited to derive sea-level records for the last 470 k.y. (Siddall et al., 2003, 2004).

Our new high-resolution Red Sea δ^{18}O record shows a clear reversal of 1.5‰ within the TII trend of decreasing δ^{18}O values (Fig. 1), while the total glacial to interglacial δ^{18}O change in Red Sea records exceeds 5‰ (Hemleben et al., 1996). Full details of the methods used to derive the Red Sea δ^{18}O record shown here are given by Hemleben et al. (1996). We note that the 1.5‰ amplitude of the mid-TII δ^{18}O reversal in the Red Sea equals that of the full glacial to interglacial δ^{18}O range observed in the adjacent Gulf of Aden (Almogi-Labin et al., 2000). Hence, the Red Sea δ^{18}O reversal cannot be explained in terms of changes in

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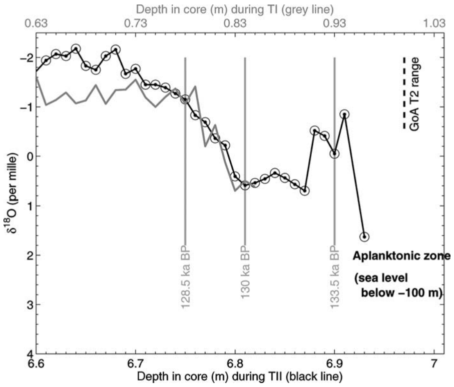

Figure 1. Central Red Sea $\delta^{18}O$ for *Globigerinoides ruber* from core GeoTue KL11 (18°44.5N, 39°20.6E) (black line). Gray line is same $\delta^{18}O$ record for Termination I (TI) from GeoTue KL11 with shifted core depth scale for comparison (top axis). Vertical gray lines indicate coral U-Th dates used as chronological tie points (Stirling et al., 1998; Esat et al., 1999; Gallup et al., 2002; Thompson and Goldstein, 2005) for ages shown in Figure 2. There is an uncertainty of ±1 k.y. in assignment of ages to early sea-level highstand and intermediate lowstand due to small number of coral age estimates for these events. Vertical dashed black line indicates full glacial to interglacial range during TII in $\delta^{18}O$ from *Globigerinoides ruber* in Gulf of Aden (GoA) core GeoTue KL15 (12°51.5N, 47°25.9E) as discussed in text (Almogi-Labin et al., 2000). $\delta^{18}O$ data are derived from RV *Meteor* cruise 5, leg 3 and are deposited in Institute of Geosciences (Bohrkernlager), University of Tübingen, Germany. Isotope values were measured at Leibnitz Laboratory, Kiel, following standard procedures. Details are given by Hemleben et al. (1996).

temperature or precipitation/evaporation associated with a regional (Younger Dryas style) climate reversal. This would imply that the reversal in the Red Sea $\delta^{18}O$ record is a locally imparted signal, which (given the magnitude) likely would reflect a significant sea-level fluctuation.

The magnitude of the Younger Dryas temperature reversal during TI was 2 °C in the northern Red Sea (Arz et al., 2003). The KL11 core site in the central Red Sea is ~1000 km away from the core site of Arz et al. (2003), and there is no evidence that temperature changes observed in the northern basin are similar to those in the central basin. Indeed Arz et al. (2003) noted important climate changes strictly limited to the northern Red Sea basin during the Holocene. Nevertheless, we consider the effect of temperature variability of 2 °C magnitude on our results. The confidence limits given for the Red Sea sea-level calibration account for ±2 °C changes in temperature as well as ±40% change in evaporation, ±10% in relative humidity, and a 0.1‰ measuring uncertainty in $\delta^{18}O$, resulting in a total 2σ sea-level uncertainty of ±12 m (Siddall et al., 2003, 2004). Variability in the modeled sea level with an amplitude greater than 12 m can therefore not be interpreted in terms of a Younger Dryas–style temperature event. The Red Sea record of the TII sea-level reversal documents a 30 ± 12 m fall in sea level following an early highstand. We note that the maximum possible peak to trough change in the Red Sea might be as large as 40 ± 12 m, but this relies on only one point during the initial highstand.

The Red Sea record of the TII sea-level reversal is placed in a strong stratigraphic framework, following the marine oxygen isotope stage (MIS) 6 lowstand and preceding the MIS 5e highstand (Fig. 1). The MIS 6 lowstand is characterized by an aplanktonic zone in cores from the central and northern Red Sea (Fenton et al., 2000); during periods of sea level much lower than −100 m, central and northern Red sea salinity reached intolerable levels for planktonic foraminifera, which became extinct for brief ''aplanktonic'' periods. The early highstand is constrained by four points with low $\delta^{18}O$ values, which are consistent with an early highstand.

Age Model

The Red Sea provides a strong stratigraphic framework for the sea-level changes suggested from observations at a number of different sites during TII. However, there is currently no way to date Red Sea cores directly during TII and thereby generate an independent Red Sea age model. Here we use information from U-Th dated coral reefs to create an age model for the Red Sea sea-level curve during TII.

The MIS 5e highstand has been well described by Stirling et al. (1998) using U-Th dated corals from Western Australia. These workers found that the earliest MIS 5e corals at this site gave U-Th ages of 128 ± 1 ka. Recently, Thompson and Goldstein (2005) revised the age estimates using a model to correct for the effects of open-system behavior on U-Th ages from a range of sources. We take 128.5 ka as one tie point for our age model to allow for the corrected open-system ages of Thomson and Goldstein (2005). This is in agreement with the Stirling et al. (1998) date of 128 ± 1 ka for the onset of MIS 5e.

Geochronological results from Aladdin's Cave corals show some indications that the samples suffered from diagenesis and that U-Th ages may be slightly inaccurate. For example, the large Porites coral found in the cave exhibited variable $\delta^{234}U_i$ values ranging from 142‰ to 164‰. Stirling et al. (1998) recommended that values outside the 145‰–153‰ window should be eliminated. In addition, even after exclusion of three Aladdin's Cave corals that were clearly deposited at a later stage, there is still a large scatter of the measured U-Th ages, which range between 126 and 134 ka, i.e., as old as the TII highstand and as young as the early phase of the MIS 5e highstand. For the sake of simplicity, we will keep the median age of 130 ka for the end of the TII lowstand as proposed by Esat et al. (1999), but this overall uncertainty should be kept in mind when making comparisons with other studies and when evaluating rates of sea-level change.

Corals used to generate age estimates for the brief TII sea-level peak prior to the reversal originate from the western part of Barbados (Gallup et al., 2002). The corals document an early growth phase corresponding to the initial TII highstand at 135.8 ± 0.8 ka (or 135.5 ka using the same decay constants as those applied by Esat et al., 1999). This is within the error of the dates presented by Esat et al. (1999). In support of an early highstand at around 133.5 ka, Stirling et al. (1998) described evidence from three drill-core observations of fossil coral reefs from Western Australia and estimated sea levels ~14 m below present (corrected for isostasy) at around 134 ka. These ages are uncorrected for open-system effects and lie at the limit or slightly outside of the acceptable range for closed-system U-Th age techniques and so are not mentioned further in the text. We mention them here because of the reproducibility of their ages in three instances. We assume an early highstand mid-point of 133.5 ka using the Thompson and Goldstein (2005) revised estimates of the U-Th ages provided by Gallup et al. (2002).

Gallup et al. (2002) found some evidence for a drop in sea level following the early highstand, but could not confirm the magnitude of the subsequent lowstand. At present uplift rates, any fossil reef from the intervening lowstand on Barbados would be below modern sea level, which may explain the lack of coral-based evidence for the magnitude of the lowstand from Barbados. Thompson and Goldstein, (2005) pointed out that sea-level constraints from submerged Bahamas speleothems (Richards et al., 1994) indicate a sea-level reduction greater than 30 m for the intervening lowstand at around 130 ka, which is in agreement with the Aladdin's Cave data.

Using the U-Th dated corals from Barbados, Aladdin's Cave, Western Australia, and the corrections for open-system behavior described by Thompson and Goldstein (2005), as well as evidence from Bahamas speleothems, we ascribe an age model to the Red Sea record. A typical rate of sea-level fall between 1 and 4 cm yr^{-1} is suggested for the sea-level reversal, similar to values of 2 cm yr^{-1} reported for episodes within MIS 3 (Siddall et al., 2003). This value is based on the age model for the sea-level curve shown in

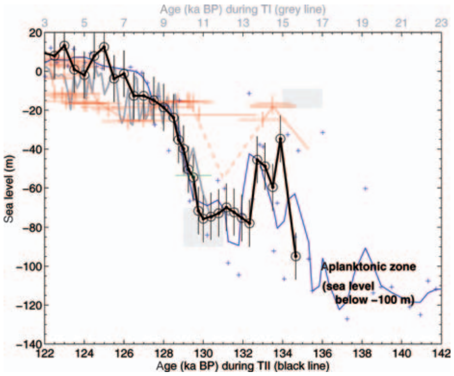

Figure 2. Red Sea sea-level estimates (thick black line and dots) with error margin (±12 m, thin black lines) along with published sea-level estimates from Esat et al. (1999) (gray boxes). Red crosses and lines are corrected U-Th ages after Thompson and Goldstein (2005) and the sea-level curve derived from them. Error bars on Thompson and Goldstein sea-level estimates (2005) allow for uncertainty in Barbados uplift rates (Schellman and Radtke, 2004). Dashed line is point where Thompson and Goldstein (2005) curve is constrained by speleothem constraints. Green cross represent dates for which Bahamas speleothems were undergoing a growth phase and error for these dates from Richards et al. (1994) (sea-level curve should be below these levels). Thick gray line is sea level calculated from $\delta^{18}O$ record for Termination I (TI) from GeoTue KL11 after Siddall et al. (2003) with shifted, age scale for comparison (top axis). Blue line is scaling to sea level of the V19–30 benthic $\delta^{18}O$ (*Uvigerina*) record after Cutler et al. (2003) with 1 k.y. added to age scale to aid comparison (5-point digitally filtered using Butterworth digital filter available in MatLab programming package—result is not sensitive to filter type. Original data, shown as blue crosses, are available at: http://delphi.esc.cam.ac.uk/coredata/wwwcoredata/VEMA/v1930uv.html).

Figure 2 as described in Figure 1 and allowing for a ±1 k.y. uncertainty in the assignment of ages to the Red Sea record (i.e., shifting assignment of coral ages to the Red Sea record with respect to core depth for the early highstand, intervening lowstand of the sea-level reversal, and the MIS 5e inception). We further note from Figure 2 the general similarity in the rate of sea-level rise and Red Sea $\delta^{18}O$ change in the latter parts of TI and TII.

INTERPRETATION AND DISCUSSION

There are general similarities between the sea-level reconstructions from the Red Sea record and coral data (Fig. 2)—all the records presented here show an early highstand followed by a lowstand. There are also some clear differences—the coral data indicate higher sea levels during the early highstand than the Red Sea reconstruction, which may be accounted for by variations in isostatic uplift and uncertainty in tectonic uplift between the sites.

The Barbados corals were taken from an area affected by the Clermont-Nose anticline (Gallup et al., 2002). The area of the Clermont Nose anticline is subject to a large spatial differential in uplift rate (Schellmann and Radtke, 2004). Such a large spatial differential in uplift in the vicinity of an anticline feature suggests that the uplift may have also varied over time (Schellmann and Radtke, 2004). It is possible that the Gallup et al. (2002) Barbados coral data may not give accurate absolute sea-level estimates for the period but nonetheless constrain the age of the initial highstand. Although similar uncertainty in the spatially varying uplift rate along Huon Peninsula and isostatic corrections prevent the derivation of accurate absolute sea-level estimates for the TII reversal, a relative change in sea level has been estimated to be between 55 and 75 m (Esat et al., 1999). This is much larger than our estimate of 30 ± 12 m or even 40 ± 12 m, which relies on the single point describing the maximum highstand. Siddall et al. (2004) discussed isostatic effects on Hanish Sill (up to 17 m lowering of the sill at the Last Glacial Maximum) and how the Red Sea method for deriving sea level may allow for these effects to some extent through a linear tuning of their model. Antonioli et al. (2004) noted that sea level estimated by the Red Sea method is lower than coral-based estimates during MIS 5a and MIS 5c. Therefore, isostatic changes to the level of Hanish Sill cannot be ruled out.

Benthic oxygen isotope records from the open ocean have been used to estimate continuous records of sea level (Shackleton, 1987; Labeyrie et al., 1987; Walebroeck et al., 2002; Cutler et al., 2003). The benthic oxygen isotope signal varies in magnitude and sensitivity to ice volume between ocean basins (Labeyrie et al., 1987). The deep Pacific has commonly been regarded as most sensitive to ice-volume changes, while the Atlantic is most sensitive to the realignment of water masses during the glacial cycle (Labeyrie et al., 1987; Skinner et al., 2003). Cutler et al. (2003) applied a simple scaling of the high-resolution Pacific V19–30 benthic isotope record (after Shackleton, 1987) to sparse, coral-derived estimates of sea level. This scaled record is shown in Figure 2 alongside the Red Sea–derived estimate. A clear reversal signal is present in the V19-30 record despite noise in the record during TII. There is good agreement between the Red Sea estimate and the scaled V19-30 record after Cutler et al. (2003) on the magnitude of the reversal.

Studies using the most direct indicators of sea level available have consistently indicated that sea level does not follow an archetypal sawtooth pattern but an irregular, stepped pattern with considerable variability (e.g., Lambeck and Chappell, 2001, and references therein). Abrupt increases in ice volume occurred both at the glacial inception and during glacial periods, suggesting that rapid growth of large continental ice sheets caused sea-level lowering at rates of 1–2 cm yr^{-1} (Cutler et al., 2003; Chappell, 2002; Siddall et al., 2003; Thompson and Goldstein, 2005). The TII sea-level reversal was therefore likely not an isolated event in terms of the rates or magnitude of sea-level change suggested here, but part of a general pattern of rapid changes in sea level observed throughout the last glacial cycles. The millennial time scale of the sea-level reversal makes for compelling parallels with other periods of millennial-scale change thought to result from internal variability within the Earth system, rather than as a response to external forcing (Oppo et al., 1998; Siddall et al., 2003; Knutti et al., 2004). We speculate that there is a link between the early sea-level peak of TII and millennial-scale internal variability within the climate system. Indeed, Thompson and Goldstein (2005) and Siddall et al. (2003) have provided evidence that millennial-scale variability in sea level is endemic in the climate system irrespective of changes in insolation.

Future work on the TII sea-level reversal should include deriving sea-surface temperatures for the Red Sea sediment proxy data to remove remaining uncertainty about the influence of temperature in Red Sea $\delta^{18}O$ on the precise form of sea-level change during TII (i.e., the 2σ uncertainty interval on sea-level reconstruction can be somewhat reduced by inclusion of independent sea-surface temperature estimates). The early highstand is not constrained by more $\delta^{18}O$ values because of the short event duration and the resolution limitations imposed by continuous centimeter-scale sampling in core KL11. Additional high-resolution core samples from the central Red Sea covering the TII reversal are needed to demonstrate the reproducibility of our result. More coral data should be sought for the early highstand and intervening lowstand, particularly since relative stillstands of several millennia duration (as indicated here for the initial highstand and intervening lowstand) are long enough to have permitted substantial coral growth. Further speleothem evidence should provide an enhanced and datable stratigraphic framework for the reversal.

CONCLUSIONS

The Red Sea $\delta^{18}O$ record during TII is explained by a reversal in the rising sea-level trend. The reversal in the Red Sea $\delta^{18}O$ record cannot be explained by temperature, evaporation, or any effects other than sea-level change. This reversal has been documented in other geological archives, and it is likely that it was a real feature of TII. Our record suggests that the initial sea-level highstand lasted several millennia and was followed by a decrease in sea level of 30 ± 12 to 40 ± 12 m, a stillstand that lasted several millennia, until the final sea-level rise into the MIS 5e interglacial. This last part of TII shows strong similarities to the last part of TI.

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