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Climate of the Past Discussions

Late Pliocene age control and composite depths at ODP Site 982, revisited

N. Khélifi[1,*], M. Sarnthein[1], and B. D. A. Naafs[2,3]

[1]Institute of Geosciences, University of Kiel, 24118 Kiel, Germany
[2]Alfred Wegener Institute for Polar and Marine Research, 27568 Bremerhaven, Germany
[3]Leibniz Center for Earth Surface and Climate Studies, Institute for Geosciences Potsdam University, 14476 Potsdam, Germany
[*]current address: Leibniz Institute of Marine Sciences (IFM-GEOMAR) at the University of Kiel, 24148 Kiel, Germany

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Correspondence to: N. Khélifi (nkhelifi@ifm-geomar.de)

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Abstract

Ocean Drilling Program (ODP) Site 982 provided a key sediment section at Rockall Plateau for reconstructing northeast Atlantic paleoceanography and monitoring benthic δ^{18}O stratigraphy over the Late Pliocene to Quaternary onset of major Northern Hemisphere Glaciation. A renewed hole-specific inspection of magnetostratigraphic events and the addition of epibenthic δ^{18}O records for short Pliocene sections in holes 982A, B, and C, crossing core breaks in the δ^{18}O record published for Hole 982B, now imply a major revision of composite core depths. After tuning to the orbitally tuned reference record LR04 the new composite δ^{18}O record results in a hiatus, where the Kaena magnetic event might been lost, and in a significant age reduction for all proxy records by 130 to 20 ka over the time span 3.2–2.7 million yr ago (Ma). Our study demonstrates the significance of reliable composite-depth scales and δ^{18}O stratigraphies in ODP sediment records for ocean-wide correlations in paleoceanography and makes Late Pliocene trends found at Site 982 much better comparable to those published from elsewhere in the North Atlantic.

1 Introduction

Several studies used the Pliocene sediment section of Ocean Drilling Program (ODP) Site 982 (Rockall Plateau; 57°3′ N, 15°5′ W; 1134 m water depth) in the North Atlantic for paleoceanographic reconstructions. In particular, Lisiecki and Raymo (2005) used the orbitally tuned δ^{18}O record of Hole 982B for establishing the LR04 stacked record, Lawrence et al. (2009) for generating a Pliocene sea surface temperature (SST) record, and Pagani et al. (2010) for reconstructing a late Pliocene pCO$_2$ drop.

The Leg 162 Shipboard Scientific Party (1996) first established an age control at Site 982 for the Upper Pliocene (3.65–2.60 Ma) by means of both magnetic reversals in Hole A (with regard to the Matuyama-Gauss boundary; Channell and Lehmann,

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1999; Channell and Guyodo, 2004) and an epibenthic $\delta^{18}O$ record mostly obtained from Hole B (Venz and Hodell, 2002).

In his PhD thesis Khélifi (2010) supplemented the epibenthic $\delta^{18}O$ record of Hole 982B and added pieces of records from Holes 982A and C in order to bridge more accurately various coring gaps (Table S1; see Supplement). On this way we identified and had to insert small additional core sections previously overlooked by the Shipboard Scientific Party (1996) in the composite depth scheme. Consequently, we had to revise the Late Pliocene chronostratigraphy at Site 982.

2 Composite depth record

The original composite-depth scale at Site 982 was derived from sediment records of continuous multi sensor logging (GRAPE density, magnetic susceptibility, *p*-wave velocity) in four holes 982A–D, moreover, from records of spectral color reflectance (Shipboard Scientific Party, 1996). However, below 54 m composite depth (m c.d.) color reflectance was only measured in holes A and B. Between 55 and 72 m c.d. all core fits amongst holes A, B, and C are poorly established as they cannot rely on specific oscillations and structures in the color reflectance and magnetic susceptibility records, here dropping to background levels. Also, poor structures in the GRAPE record appear unsatisfactory for composing a composite depth model (see Shipboard Scientific Party, 1996, Fig. 2 in Chapter 4).

Thus, most correlations between holes A, B, and C are hardly unique but speculative, in particular amongst cores B5 to B8, A6–A8, and C7–C10. In contrast, the oscillations of overlapping benthic $\delta^{18}O$ records of holes 982A–C show numerous marked and unique structures that appear suitable for establishing a reliable continuous composite depth scale (Fig. 1).

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3 Magnetostratigraphy framework

The magnetic polarity of the upper Gauss chronozone (2An) is well reflected at Site 982 by core sections with normal inclination (Channell and Lehmann, 1999; Channell and Guyodo, 2004). In Hole A it ends with a sharp change in magnetic polarity at 57.23 m c.d. (50.70 m b.s.f.) (Fig. 1). This depth of the Gauss/Matuyama (G/M) boundary was by now employed by all authors since Venz and Hodell (2002) for age-calibrating marine isotope stage (MIS) 104 in the Hole-B δ^{18}O record, where they had bridged the core break B6–B7 by a short piece of δ^{18}O record from Hole A (57.5–59.69 m c.d.). However, this position of the G/M boundary at 57.23 m c.d. (dated at ~2.608 Ma in LR04) is different from that found in Hole B, where the midpoint of the reversal occurs at 58.06 m c.d. (~52.55 m b.s.f.), ~0.8 m deeper.

In this study we take care of this discrepancy, in part, by measuring an additional short δ^{18}O record over the crucial core section in Hole B between 57.51 and 59.67 m c.d. (Fig. 1). We are aware of a number of subsequent, very short-term oscillations toward positive magnetic polarity, that occur further upcore, up to 54.81 m c.d., and are as yet unexplained. In harmony with Channell and Lehmann (1999) we lump these excursions with the Matuyama chron (Fig. 1).

4 Stable isotope stratigraphy and composite depths

The initial Pliocene δ^{18}O record from Site 982 of Venz and Hodell (2002) was slightly supplemented by Lisiecki and Raymo (2005). These authors already identified a short hiatus between 2.25 and 2.33 Ma that removed MIS 86, 87, and 88. This gap is important in demonstrating that pelagic sedimentation at Site 982 was occasionally discontinuous.

For the present study we took ~345 sediment samples from holes 982A, B, and C at the Bremen Integrated Ocean Drilling Program (IODP) Core Repository (Table S1; see Supplement). The samples were weighed, oven-dried at 40 °C, weighed again (to obtain dry bulk density), and washed over a 63-µm mesh-size sieve. The residue was dried at 40 °C and finally sieved into 5 size fractions. *Cibicidoides wuellerstorfi* and *C. mundulus* were picked from the >250-µm fraction and analyzed for stable isotopes on a Finnigan MAT251 system, with a precision of ±0.07 ‰ for $\delta^{18}O$ and ±0.05 ‰ for $\delta^{13}C$, at the Leibniz Laboratory in Kiel. $\delta^{18}O$ values of *Cibicidoides* spp. were corrected by 0.64 ‰ (Ganssen, 1983) (Table S1; see Supplement) to normalize them to $\delta^{18}O$ values of Uvigerina, which are in equilibrium with ambient seawater (Shackleton, 1974).

The glacial-to-interglacial $\delta^{18}O$ oscillations of our new benthic $\delta^{18}O$ records now produce a series of new splice tie points between holes 982A, B, and C, which suggest a revision of the former splice of Hole-A, Hole-B, and Hole-C core sections at 63.68–~85.00 m c.d. (Fig. 1b) (Table S2; see Supplement), initially defined by the Shipboard Scientific Party (1996) (Fig. 1d). Our new tie points also modify the chronostratigraphic scheme proposed by Venz and Hodell (2002), which was tuned and incorporated into LR04 by Lisiecki and Raymo (2005). Admittedly, also the new core splice displays some minor, as yet unexplained deviations in core fit. At 66.37–68.36 m c.d. (MIS G19–20) the absolute $\delta^{18}O$ levels in holes B and C are up to >0.3 ‰ higher than in Hole A. On the other hand, the $\delta^{18}O$ level in Hole B is up to 0.3 ‰ lower than in holes A and C at ~70.31–70.76 (new) m c.d. (MIS KM5). In general, however, the intercore deviations of $\delta^{18}O$ excursions remain within the range of analytical uncertainty (±0.07 ‰; Fig. 1b).

Below 68.36 m c.d. the new splice of composite depths requires an additional downhole shift of the core top B8 by ~0.5 m. By comparison with holes A and C core break B7–8 now covers a gap of ~2 m, from 68.36 to at least ~70.31 m c.d., whereas core break B8–9 has lost a section between ~79.4 and 80.5 m c.d. (Fig. 1b vs. d).

5 Discussion of correlations in stable-isotope stratigraphy

At Site 982, the published definitions of Late Pliocene MIS mostly relied on the largely continuous benthic $\delta^{18}O$ record of Hole B. Short $\delta^{18}O$ records were measured in Hole A for bridging core breaks B6–7, B7–8, and B8–9 (Venz and Hodell, 2002;

supplemented by Lisiecki and Raymo, 2005). We now prolonged the δ^{18}O master record of Hole B for short sections at the base of core B6 and top of B7 between 57.51 and 59.67 m c.d. and replaced the G/M boundary in Hole A at 50.70 m b.s.f. (57.23 m c.d.) by that in Hole B at 52.55 m b.s.f. (58.06 m c.d.) to be consistent with age calibrating MIS 104 now defined in Hole B. Accordingly, the composite depth position of the G/M boundary shifted 0.83 m downcore. This shift also implied a redefinition of the numbers of several preceding MIS.

Figure 2 shows the resulting new suite of benthic δ^{18}O oscillations in holes A, B, and C, re-adjusted to the orbitally tuned reference record LR04. Different from LR04, MIS 104 appears bipartite, with a second, very short cold excursion in the upper part, a structure likewise found in the δ^{18}O records of holes A and B. However, a bipartite benthic MIS 104 is also found at nearby Site 548 at similar water depths in the northeast Atlantic, a site that likewise was influenced by Mediterranean-source bottom waters (Khélifi et al., 2009).

Subsequent to MIS 104, the suite of MIS 103–100 closely reflects all details of amplitude and structure as defined in LR04 and reported in previous studies. The influence of Mediterranean waters was obviously too weak for changing significantly the benthic δ^{18}O signal at Site 982 (Khélifi, 2010).

However, different from previous studies, all stages of the suite G1 to G10 prior to MIS 104 (redefined at 58.06 m c.d.) now resemble in detail the structures and amplitudes displayed in LR04. Perhaps, the δ^{18}O excursion at glacial MIS G4 is a little less pronounced than the coeval signal in LR04 (Fig. 2). However, once more it is analogous to the δ^{18}O record at Site 548 (Khélifi et al., 2009). In particular, the amplitude of G10 now clearly exceeds that of all preceding "cold" MIS, as suggested by LR04, thus strongly differs from the previous tuned version. Further back, from MIS G11 (~2.83 Ma) back to G20 (~3.01 Ma), the redefined MIS oscillations likewise resemble closely those of LR04. MIS G13 and in particular MIS G15 depict bipartite peaks as suggested by LR04.

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In contrast to the outlined general match of MIS structures the amplitude of MIS G13 appears clearly "warmer" in Hole B than in Hole A and LR04. In contrast to LR04 the high amplitude of MIS G13 exceeds that of bipartite MIS G17. However, G13–G17 show a similar suite of amplitudes in ODP Site 846 (Tiedemann et al., 1994). MIS G13 coincides with a very short, marked interval of questionable negative inclination within the Gauss Chron near 63.72 m c.d. in Hole B (depicted in Fig. 1a and b despite of low magnetization intensities; Chanell and Lehman, 1999). The "excursion" cannot be correlated to the theoretically nearby Kaena event, and thus is discarded. Further downcore, the amplitudes of MIS G19 in holes B and C are as high as suggested by LR04, that is lower than the amplitude of G17. In contrast, the relative amplitude of MIS 19 in Hole A clearly exceeds that of LR04. The broad and marked positive δ^{18}O excursion of MIS 20 in holes B and C is somewhat more distinct and thus comes much closer to that suggested by LR04 than the weak G20 excursion found in Hole A (Fig. 2). We do not understand yet the more negative δ^{18}O levels in the Hole-A record, although the relative amplitudes of the δ^{18}O excursions for G20 are largely the same in all three holes.

Prior to MIS G20, that is now the lower end of the δ^{18}O record in core B7, the relative amplitudes of MIS G21 and G22 in the (new) Hole A and C records now match closely the pertinent features in LR04. However, these two records do not extend beyond the base of MIS G22 and do not reach back to the δ^{18}O spike at the top of core B8, which probably occurs right at the end of KM5. Thus the joint δ^{18}O record of the continuous sediment sections in holes A and C do not bridge completely the (now expanded) coring gap at core break B7–B8 between 68.36 and ~70.31 m c.d. (Figs. 1b and 2). A correlation with LR04 suggests that the newly defined data-devoid interval from top K1 back to KM4 is interpreted as major stratigraphic hiatus, per analogy with the gap found for MIS 86, 87, and 88. The loss of MIS K1–K2 and KM1–KM4 necessarily would also imply a complete loss of the magnetic signals that traced the coeval Kaena magnetic event (poorly documented anyway because of low magnetization intensities). We are aware that these shifts still leave unsolved a discrepancy between the δ^{18}O record of

Hole A and the parallel records of holes B and C at stage KM5, possibly a result of small-scale differences in local sediment loss.

Prior to KM5 the records of holes A and C require only minor shifts in composite depth near MG2–MG5 to adjust them closely to the MIS record of Hole B (Table S2; see Supplement). The stage chronology of Hole B is retained unchanged back to MIS Gi2 such as previously tuned to LR04 by Lisiecki and Raymo (2005).

Near 74.19 (new) m c.d., the onset of a significant positive excursion in δ^{18}O marks the onset of the extremely cold MIS M2, which can be aligned with a well-resolved event of negative inclination within the Gauss Chron (Channell and Lehmann, 1999) (Fig. 1a). Accordingly, we may now consider this magnetic excursion as part of the Mammoth subchron at ~3.3 Ma (Lisiecki and Raymo, 2005). The upper part of this subchron was probably lost. Unfortunately, the low magnetization intensities do not allow a proper identification of the Gauss/Gilbert boundary.

6 Implications

Our new stratigraphic correlations imply a revised chronology for the $U^{k'}_{37}$-based SST record of Site 982, that covers the Late Pliocene onset of major Northern Hemisphere Glaciation (Lawrence et al., 2009). Subsequent to 3.2 Ma we need to allow for major age reductions that vary between more than 130 and 20 ka, because the age of a δ^{18}O signal formerly assigned to MIS KM4 (~3.175 Ma) is now replaced by the age of MIS G22 (~3.045 Ma) (Fig. 3).

Accordingly, we see three Late Pliocene main cooling events of northeast Atlantic SST, which Lawrence et al. (2009) constrained to ~4.5 °C each. Apart from a first and short-term reversible cooling linked to MIS M2, the first major long-term cooling now shifted to ~3.24–3.02 Ma, that includes the time of the hiatus, the second cooling to ~2.7–2.53 Ma. An intermediate phase of “asymptotical” warming by 3 °C lasted from ~2.98 until 2.7 Ma. On the basis of our revised chronology these ages and SST trends now compare in detail with the ages and trends of various SST changes found farther north in the northern North Atlantic (Bartoli et al., 2005).

Likewise our new age model suggests that the most of a major Late Pliocene reduction in atmospheric CO_2, the start of which was placed at Site 982 somewhere near 3.2 Ma (Pagani et al., 2010), is now shifting by about 130 ky to the interval after MIS G21. This timing is in harmony with that of decreasing bottom water temperatures in the North Atlantic, then tracing a change in meridional overturning circulation (Sarnthein et al., 2009).

7 Conclusions

In our study, we revised the age-calibration of the Late Pliocene δ^{18}O master record measured in Hole 982B. In contrast to previous schemes of age control (Venz and Hodell, 2002; Lisiecki and Raymo, 2005; Lawrence et al., 2009), we replaced – here without changing the composite-depth scale published by the Shipboard Scienific Party (1996) – the G/M boundary in Hole 982A by that in Hole 982B (Channell and Lehmann, 1999), which occurs ~0.83 m farther downhole and corresponds directly to MIS 104 in the Hole-B δ^{18}O record. Moreover, the δ^{18}O record of Hole 982B near to the core break B6–B7 was prolonged farther up- and downcore. In total, these changes led to a reduction of sediment ages by up to 80 ka and a significantly improved match of MIS G1–G12 structures with those found in the reference record LR04.

New δ^{18}O records from holes 982A and C at ~64–84 m c.d., measured in parallel to the master record of Hole B, moreover, many additional data to improve on the resolution of the Hole-B record helped us to improve on the definition of stage numbers around core breaks B7–B8 and B8–B9. Here, the optimum mutual overlap amongst all three δ^{18}O records and their closest-possible match with MIS structures defined in LR04 also require a major revision of the composite-depth scale displayed in Table S2 (see Supplement). Most important, they suggest a natural hiatus of more than 130 ka between the base of MIS KM4 and the top of MIS K1. This gap corroborates the

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outlined age shift at MIS G22–G1. In contrast to the outlined age shifts, previous definitions of MIS KM5 back to Gi2 are confirmed. Here, it was possible to identify the Mammoth magnetic event in Hole 982B at the base of MIS M2. The large-scale age reduction subsequent to MIS KM5 implies a number of reasonable shifts in the tipping points of paleoclimatic signals that mark the Late Pliocene onset of major Northern Hemisphere glaciation.

Supplementary material related to this article is available online at: http://www.clim-past-discuss.net/7/1631/2011/cpd-7-1631-2011-supplement.pdf.

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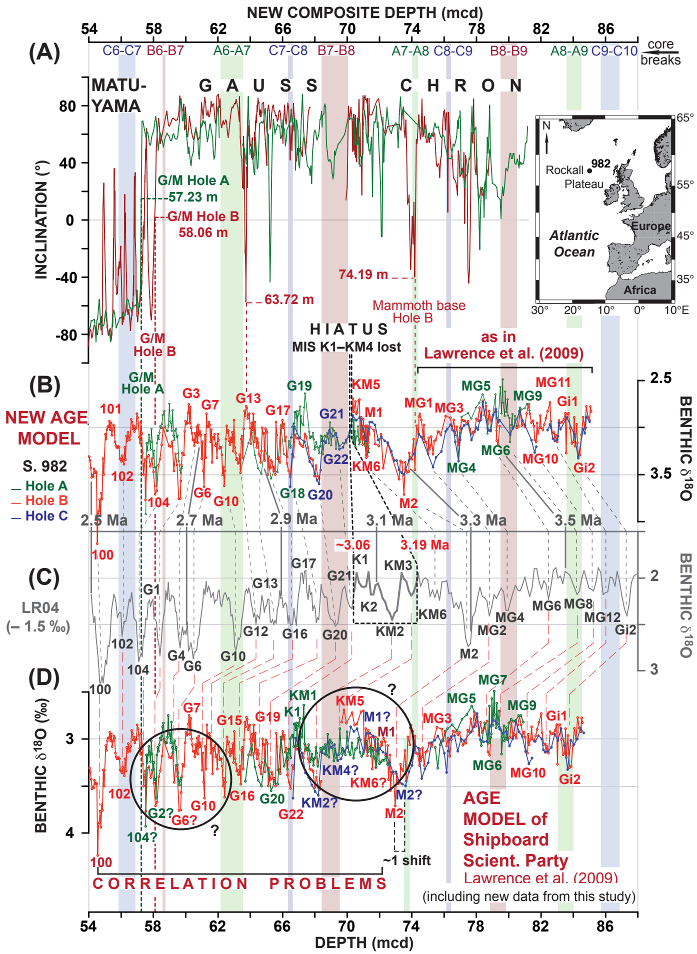

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Fig. 1. Different models of age control for ODP Site 982 (location shown at inserted map). **(A)** Magnetic polarity reversals are from Shipboard inclination data for Holes A (green) and B (red) (Channell and Lehman, 1999; Channell and Guyodo, 2004), plotted on new composite depth scale. G/M = Gauss/Matuyama. **(B)** Benthic δ^{18}O records of holes 982A–C plotted over the new composite depth scale (Venz and Hodell, 2002; supplemented by Lisiecki and Raymo, 2005, and by this study, as specified in Table S1; see Supplement). Labeled isotope stages are tuned (hatched lines) to **(C)** benthic δ^{18}O stack LR04 (Lisiecki and Raymo, 2005). **(D)** Benthic δ^{18}O records of holes 982A–C plotted over the age model of Shipboard Scientific Party (1996) as cited by Lawrence et al. (2009). Vertical bars mark potential losses of sediment sections at core breaks that differ for the old and new composite depth scales. This age model leads to major discrepancies between δ^{18}O records of Holes A, B, and C between 68 and 72 m c.d.

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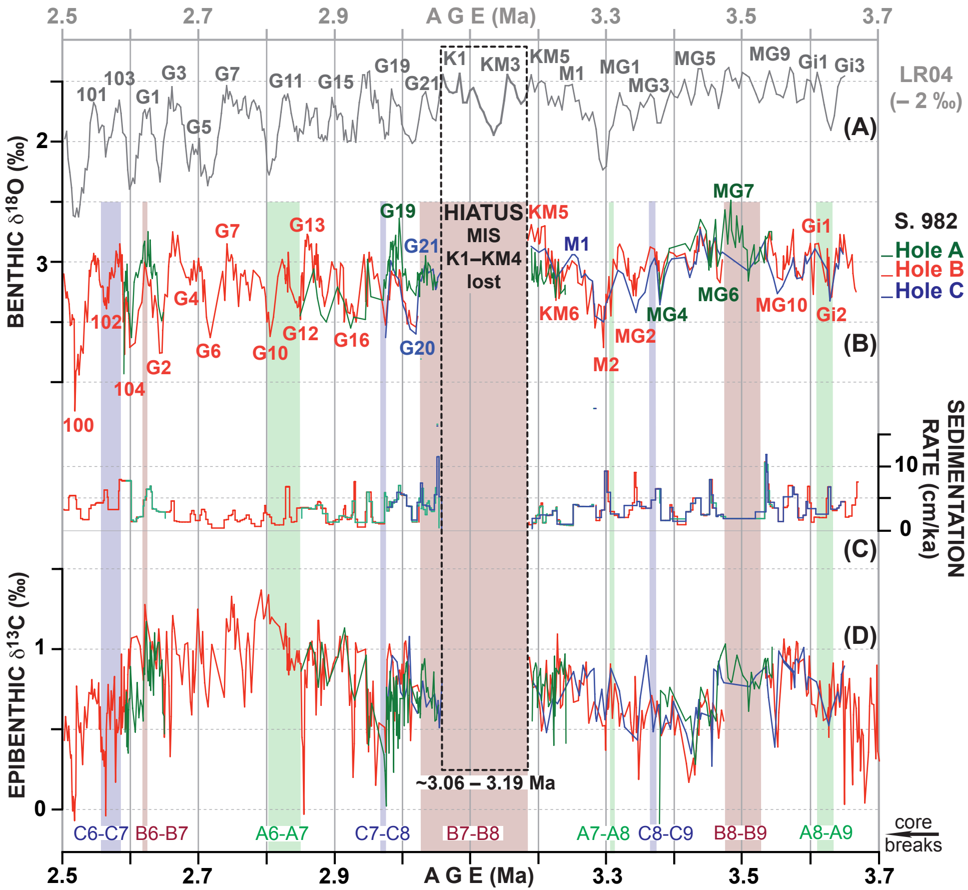

Fig. 2. Newly established δ^{18}O records and marine-isotope stage chronology **(B)** for ODP holes 982A–C versus age and LR04 ice volume record **(A)**, resulting sedimentation rates **(C)**, and composite epibenthic δ^{13}C record of holes 982A–C **(D)**.

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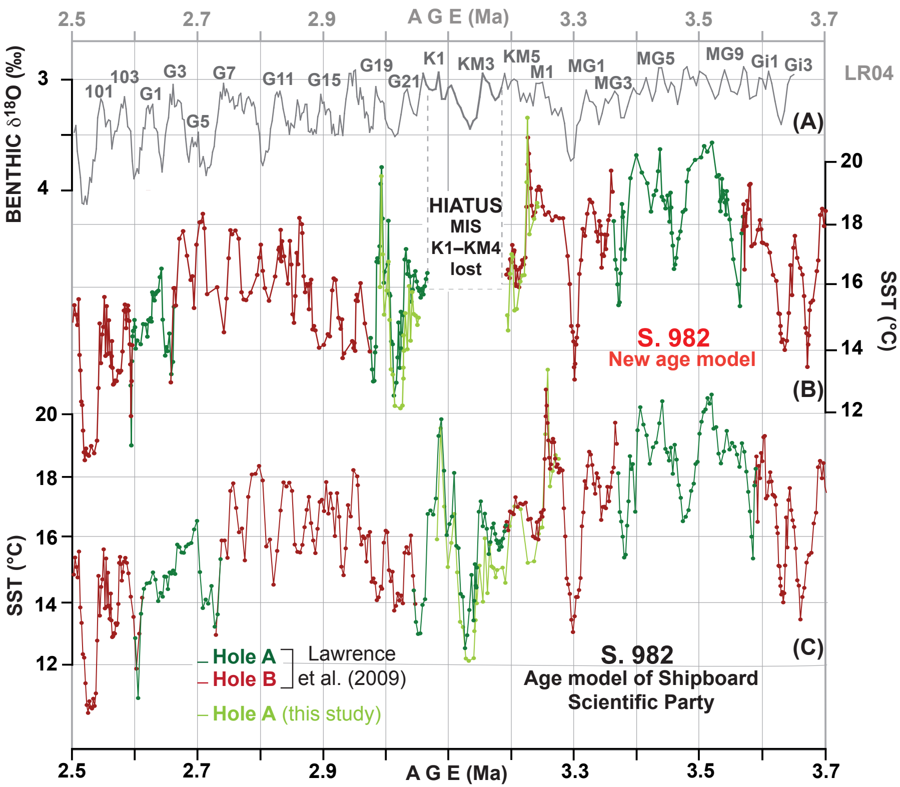

Fig. 3. $U_{37}^{k'}$-based SST record of holes A (green) and B (red) at ODP Site 982 (Lawrence et al., 2009). A LECO Pegasus III GC/TOF-MS system was used to supplement 42 $U_{37}^{k'}$ data (light green) for Hole A. This method has the advantage over the classical GC/FID methods that it reduces instrumental time and is more sensitive (further details in Hefter, 2008 and Naafs et al., 2010). Following Lawrence et al. (2009) we converted $U_{37}^{k'}$ values into SST using the calibration of Prahl et al. (1988).

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