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A Low Threshold for North Atlantic Ice Rafting from "Low-Slung Slippery" Late Pliocene Ice Sheets

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A low threshold for North Atlantic ice rafting from "low-slung slippery" late Pliocene ice sheets

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[1] Suborbital variability in late Pleistocene records of ice-rafted debris and sea surface temperature in the North Atlantic Ocean appears most extreme during times of enlarged ice sheets with a well-constrained benthic oxygen isotope-defined "ice volume threshold" ($\delta^{18}O_T$) for the "100 ka (inter)glacial" world. Information on climate instability for the earlier Pleistocene and late Pliocene is more fragmentary and/or of much lower temporal resolution, but the data available suggest similar behavior with $\delta^{18}O_T$ remaining more or less constant over the past 3000 ka. This finding is puzzling because it implies that ice rafting is highly sensitive to ice volume on short (suborbital/glacial-interglacial) time scales but not to the long-term changes in ice sheet composition associated with intensification of Northern Hemisphere glaciation (NHG). Here we report new high-resolution records of stable isotope change and ice rafting in the North Atlantic at Integrated Ocean Drilling Program Site U1308 (reoccupation of Deep Sea Drilling Project Site 609) during two glacials key to intensification of NHG (marine isotope stages G4, ~2640 ka, and 100, ~2520 ka). We find a pattern of suborbital ice rafting events showing clear evidence of threshold behavior. However, contrary to previous reports, we find that $\delta^{18}O_T$ for the late Pliocene is up to 0.45‰ Vienna Peedee belemnite (VPDB) lower than for the late Pleistocene. Using published Plio-Pleistocene global sea level records, we evaluate different potential explanations for this finding. We conclude that the observed Pliocene-Pleistocene offset in $\delta^{18}O_T$ is attributable to the existence of low-slung Pliocene ice sheets that flowed more readily than their late Pleistocene counterparts, associated with a smaller contemporaneous continental ice volume and less isotopically depleted ice.

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1. Introduction

[2] Records from the North Atlantic Ocean and Greenland ice cores show evidence of clear suborbital climate change in the late Pleistocene indicating that the ocean-atmosphere system is capable of undergoing profound reorganization extremely rapidly [*Bond et al.*, 1993, 1997; *Broecker*, 1997]. Ice sheets are widely thought to at least resonate with or amplify such change through either fresh water release to the oceans and/or regulation of sea ice extent [*Raymo et al.*, 1992; *Alley et al.*, 1999; *Wara et al.*, 2000; *Shackleton et al.*, 2000; *Mc Intyre et al.*, 2001]. Suborbital variability in late Pleistocene records appears most extreme during times of enlarged ice sheets and maximum discharge of icebergs to the North Atlantic ice-rafted debris (IRD) Belt [*Oppo et al.*, 1998; *McManus et al.*, 1999]. In fact, a benthic oxygen isotope-defined "ice volume threshold," $\delta^{18}O_T$ ($\delta^{18}O$ in *Cibicidoides wuellerstorfi* = +3.5‰ ± 0.08‰ Vienna Peedee belemnite (VPDB) at 1σ level; hereinafter $\delta^{18}O_{T\text{-}Mc99}$), has been hypothesized [*McManus et al.*, 1999] beyond which suborbital ice rafting events are instigated and sea surface temperature variability is amplified in the subpolar North Atlantic during the "100 ka (inter)glacial" world of the last five glacial cycles (Figure 1).

[3] Suborbital events in IRD flux to the North Atlantic Ocean are now also known from fragments of the early Pleistocene "41 ka (inter)glacial world" (~1940–1760 ka, marine isotope stages (MIS) 70 to 64 [*Mc Intyre et al.*, 2001]; ~1350–1275 ka, MIS 44 to 40 [*Raymo et al.*, 1998]) and these appear to indicate similar threshold-like behavior associated with near-identical benthic $\delta^{18}O$ values (+3.4 to 3.5‰ VPDB) (Figure 2a). Furthermore, long, lower-resolution records from multiple Pliocene sites have been interpreted to demonstrate that the first region-wide appearance of IRD in the subpolar North Atlantic Ocean during the intensification of Northern Hemisphere glaciation (NHG), 3500–2500 ka [*Maslin et al.*, 1998; *Jansen et al.*, 2000], is also associated with benthic $\delta^{18}O$ values surpassing +3.5‰ VPDB [*Kleiven*

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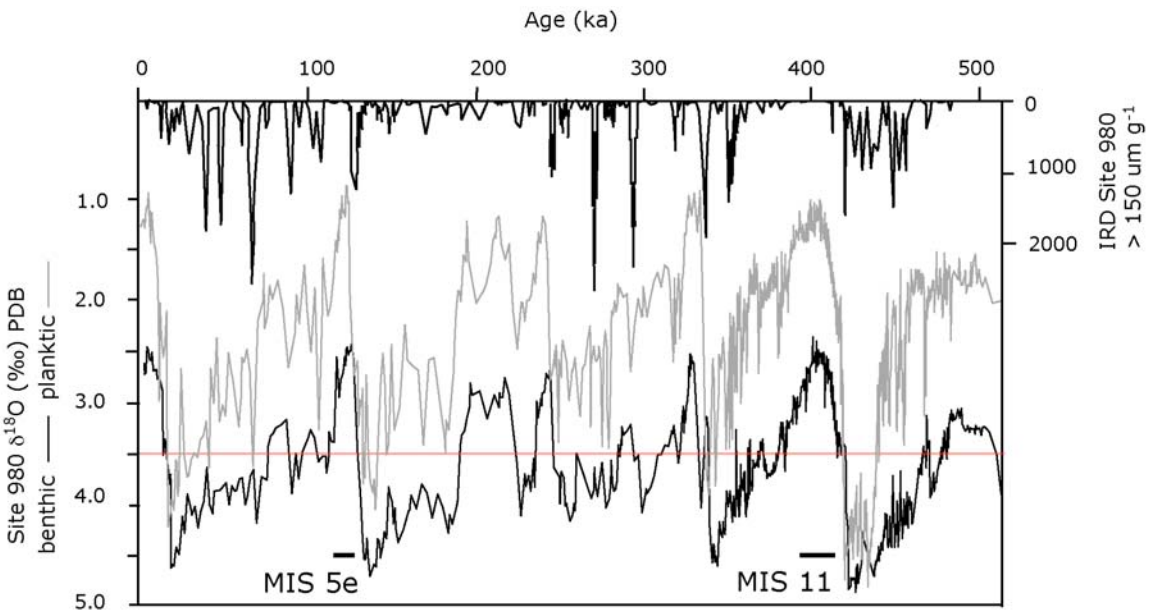

Figure 1. The relationship between ice rafting and planktic and benthic δ^{18}O (in *N. pachyderma dextral* and *C. wuellerstorfi*, respectively) for the last five glacial cycles at ODP Site 980, Feni Drift [*McManus et al.*, 1999]. Horizontal pink dashed line denotes the "threshold" benthic δ^{18}O value (*C. wuellerstorfi*, unadjusted to equilibrium; see text) associated with (1) the onset and cessation of suborbital IRD events and (2) the amplification of suborbital change in planktic δ^{18}O.

et al., 2002] (hereinafter $\delta^{18}O_{T\text{-}KI02}$, Figure 2b). In other words, it has been suggested that both the degree of climate stability at the suborbital scale and the secular onset of region-wide ice rafting in the North Atlantic Ocean are closely coupled to the growth of (Northern Hemisphere, NH) ice sheets to a threshold size, as tracked by proxy through benthic δ^{18}O ($\delta^{18}O_T \sim +3.5$‰ VPDB). If the concept of $\delta^{18}O_T$ is valid, then these findings present us with something of an enigma. On one hand the onset and cessation of suborbital fluctuations in ice rafting in the North Atlantic appear highly sensitive to global ice volume on the glacial-interglacial time scale. On the other hand, if $\delta^{18}O_T$ has remained more or less constant over the past 2000 or 3000 ka, this implies a lack of sensitivity in ice rafting to the large increase in ice budgets (and/or potential changes in ice sheet composition) associated with contemporaneous long-term intensification of NHG (but see section 4).

[4] To assess this enigma we present new suborbital resolution records (~sixty to forty data points per glacial cycle) of late Pliocene ice rafting to the central North Atlantic at Integrated Ocean Drilling Program (IODP) Site U1308 (Figure 3) for two marine isotope stages, G4 and 100 (~2640 and ~2520 ka, respectively), key to the history of intensification of NHG. We find that, during the late Pliocene, the pattern of suborbital ice rafting shows clear evidence of threshold behavior but, contrary to previous work, we find that $\delta^{18}O_T$ for these slices of late Pliocene time are up to about 0.45‰ VPDB lower than reported for the late Pleistocene. We evaluate the concept of an ice volume threshold in light of this new finding and, using previously published Plio-Pleistocene global sea level curves [*Bintanja and van de Wal*, 2008; *Rohling et al.*, 2008], simple mass balance calculations and numerical modeling experiments, consider different potential explanations for this observation. We conclude that the threshold for initiation and cessation of suborbital ice rafting events in the Pliocene North Atlantic was associated with lower-slung ice sheets than their late Pleistocene counterparts, likely associated with a smaller contemporaneous continental ice volume and/or less isotopically depleted ice. This finding is in keeping with the suggestion [*Clark and Pollard*, 1998] that Pliocene ice sheets were characterized by a lower aspect ratio than Pleistocene ones arising from their growth on soft substrates following a long precursor period of sedimentation in a less glacial weathering regime.

2. Materials and Methods

2.1. IODP Site U1308

[5] We examined sediments from IODP Site U1308 (Figure 3) located on the central eastern flank of the mid-North Atlantic ridge (present water depth ~3800 m). These sediments have some of the highest accumulation rates (~6.5–11 cm ka^{-1}) known for the late Pliocene central North Atlantic (compare to those reported by *Mudelsee and Raymo* [2005, Table 2]). IODP Site U1308 is situated within the North Atlantic IRD belt which trends approximately east-west along latitudes 45 to 50°N and denotes the zone of maximum IRD deposition and hence iceberg melting during the Last Glacial Cycle, LGC (Figure 3) [*Ruddiman*, 1977a, 1977b]. Site U1308 was drilled at the same locality as Deep Sea Drilling Project (DSDP) Site 609. DSDP Site 609 was drilled using traditional coring methods and so did not

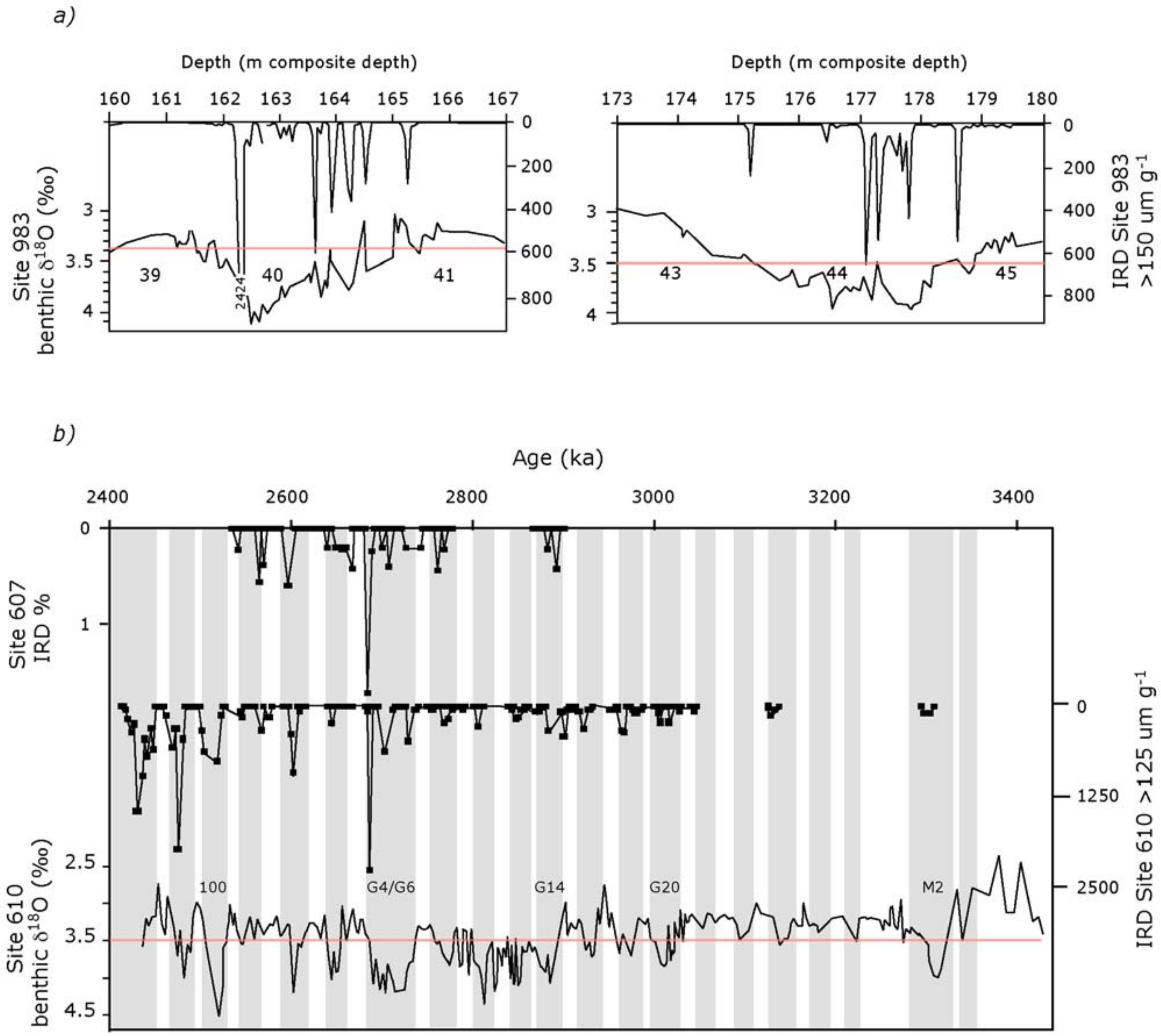

Figure 2. The relationship between ice rafting and benthic $\delta^{18}O$ for (a) the early Pleistocene, MIS 44 and 40 at ODP Site 983, Feni Drift [*Raymo et al.*, 1998] and (b) the late Pliocene at DSDP Site 610 [*Kleiven et al.*, 2002]. Horizontal pink lines denote "threshold" benthic $\delta^{18}O$ values associated with the onset of IRD deposition in the subpolar North Atlantic.

recover a composite section. Nevertheless, records from Site 609 were key to establishing the existence of Pleistocene suborbital-scale variability in the glacial North Atlantic, its correlation with Greenland ice core records and the recognition and understanding of Heinrich-events [*Bond et al.*, 1992, 1993; *Bond and Lotti*, 1995]. Reoccupation of DSDP Site 609 by IODP to recover a composite section at Site U1308 therefore presents an ideal opportunity to generate comparative records for the late Pliocene.

2.2. Stable Isotope and Ice-Rafted Debris Measurements

[6] Oxygen and carbon isotope measurements were made on foraminiferal calcite secreted by *Cibicidoides wuellerstorfi* (n = 128), supplemented where necessary by analyses of *Oridorsalis umbonatus* (n = 36), with a depth spacing of every 2 cm (at 189.45–191.23 mcd), 10 cm (at 191.33–193.79 and 198.29–200.38 mcd) and 30 cm (at 200.68–206.42 mcd) over about 17 m of the composite section to capture key intervals in the late Pliocene intensification of NHG (Figure 4). Samples (20 cc) were taken according to the shipboard primary splice [*Expedition 303 Scientists*, 2006]. Specimens were picked from the >212 μm size fraction of washed sediment samples, and typically one to four individuals were used for analysis. These were sonicated in ethanol for 5 to 10 s, oven-dried at 40°C over night, and measured on the Kiel Device I/Finnigan MAT251 system at the Leibniz Laboratory in Kiel. Analytical precision

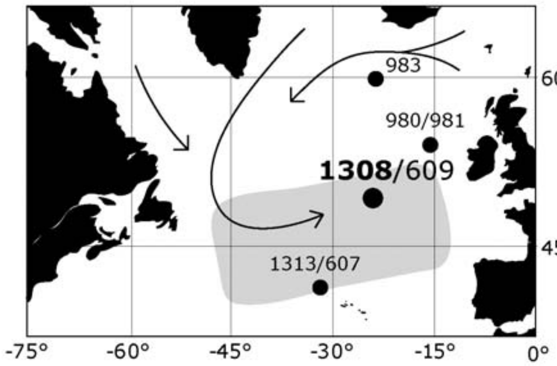

Figure 3. Location of IODP Site U1308/DSDP 609 and other sites referred to in text. Gray shaded area highlights approximate position of the Central North Atlantic Ice-rafted Debris Belt during the Last Glacial [*Ruddiman*, 1977a, 1977b]. Arrows denote simplified glacial North Atlantic "cyclonic" surface currents after *Ruddiman* [1977a, 1977b].

is ±0.07‰ (1σ) relative to the Vienna Pee Dee Belemnite, VPDB, standard for δ^{18}O and ±0.05‰ (1σ) δ^{13}C. *C. wuellerstorfi* δ^{18}O values were adjusted for species-specific offsets relative to *O. umbonatus* (widely considered to precipitate its calcite in isotopic equilibrium with respect to seawater) by adding +0.64‰ VPDB [*Shackleton and Hall*, 1984].

[7] To examine suborbital variability in ice rafting to the central North Atlantic during MIS G4 and 100 we measured the concentration of lithic grains (>150 μm g^{-1}, every 2 cm, or ~180–300 a) using a standard method [e.g., *Bond and Lotti*, 1995]. Replicate counts on some of the samples (n = 6) show a standard deviation of 1.4%. To establish the relationship between coarse lithic concentration and carbonate dissolution, we also quantified the number of whole planktic foraminifer tests and foraminifer test fragments in each sample, expressed as percent, %, foraminifer tests and the fragmentation rate of planktic foraminifer ((number of fragments/number of fragments + number of whole foraminifer tests) × 100 [*Ivanova et al.*, 2003]), respectively.

[8] Based on three lines of evidence we interpret one peak in lithics at Site U1308 during full glacial conditions of MIS 100 (lithics peak at 1900 g^{-1} across 189.99–190.13 mcd) in part to be a tephra deposit: (1) Macroscopically, this horizon is dark and massive in appearance with a sharp base and a fining upward grain size distribution. (2) Compositionally, the horizon shows a substantial increase over background levels in basaltic glass and pumice lithics. (3) Antiphasing between the relative abundance of whole foraminifer tests and foraminifer fragments (dissolution cycles) breaks down across this horizon (190–190.10 mcd) and both parameters decrease rapidly to near zero values, indicating dilution by rapid deposition of reworked volcanic material. Similarly, core photos indicate that the extremely high lithic grain counts (3500 g^{-1}, 95% quartz) during full glacial conditions in MIS G4 over three samples (198.98–199.02 mcd) are probably the result of local concentration by burrowing. Our grain counts therefore report both total lithics and lithics minus volcanics.

3. Pliocene Stratigraphy of Site U1308

[9] We focus on MIS G4 and 100 because these are two glacials key to understanding the history of ice rafting to the North Atlantic Ocean during late Pliocene intensification of NHG. MIS 100 is the earliest large-amplitude (~1.1‰) Pliocene obliquity-paced (inter)glacial cycle recorded in the benthic δ^{18}O stack (Figure 1) [*Lisiecki and Raymo*, 2005] and provides the template for our current understanding of Pliocene suborbital North Atlantic ice rafting [*Becker et al.*, 2006]. MIS G4 represents the first late Pliocene glacial where abundant IRD is documented in all subpolar North Atlantic records (Figure 2b) [*Kleiven et al.*, 2002]. Our chronology was generated by tuning our new δ^{18}O records to the LR04 global benthic oxygen isotope stack [*Lisiecki and Raymo*, 2005] in combination with the shipboard paleomagnetic reversal stratigraphy (Gauss/Matuyama boundary, 2581 ka) and lithostratigraphic data [*Expedition 303 Scientists*, 2006]. We identify a hiatus spanning the deglaciation of MIS 100 at 189.87 mcd based on a jump in benthic δ^{18}O of ~1.5‰ VPDB over a 2 cm interval. This hiatus through part of such a key glacial cycle at this locality serves to emphasize the importance of reconstructing paleorecords from continuously cored sections (the classic records from DSDP Site 609 show no warning of the presence of this hiatus because MIS 100 lies within a coring gap). However, the hiatus does not compromise our study because it falls stratigraphically above the interval capturing the descent into the glacial where the suborbital signal is known to be concentrated [e.g., *Maslin et al.*, 1995; *McManus et al.*, 1999; *Becker et al.*, 2006].

4. Confusion in the Literature Over Threshold Values in Benthic δ^{18}O

[10] To compare oxygen isotope records generated in different laboratories it is essential that data sets are reported in the same way. Even when interlaboratory analytical offsets have been accounted for and all of the isotope records under consideration have been generated on the same species it still may be necessary to adjust records where offsets from equilibrium pertain. Of particular importance here, *C. wuellerstorfi* is a species of epifaunal benthic foraminifer commonly used in paleoceanography that is widely thought to secrete calcite out of oxygen isotope equilibrium with seawater [after *Shackleton and Hall*, 1984]. In some cases *C. wuellerstorfi* δ^{18}O records are reported adjusted to equilibrium calcite (by the addition of 0.64‰ VPDB), in others they are reported unadjusted to equilibrium.

[11] The hypothesized late Pleistocene ice volume threshold of *McManus et al.* [1999] for the onset of suborbital ice rafting events and amplification of sea surface temperature variability in the subpolar North Atlantic during the last five glacial cycles ($\delta^{18}O_{T\text{-}Mc99}$) is based on unadjusted oxygen isotope ratios in *C. wuellerstorfi* (Ocean Drilling Program, ODP, Site 980). Confusingly, however, the North Atlantic

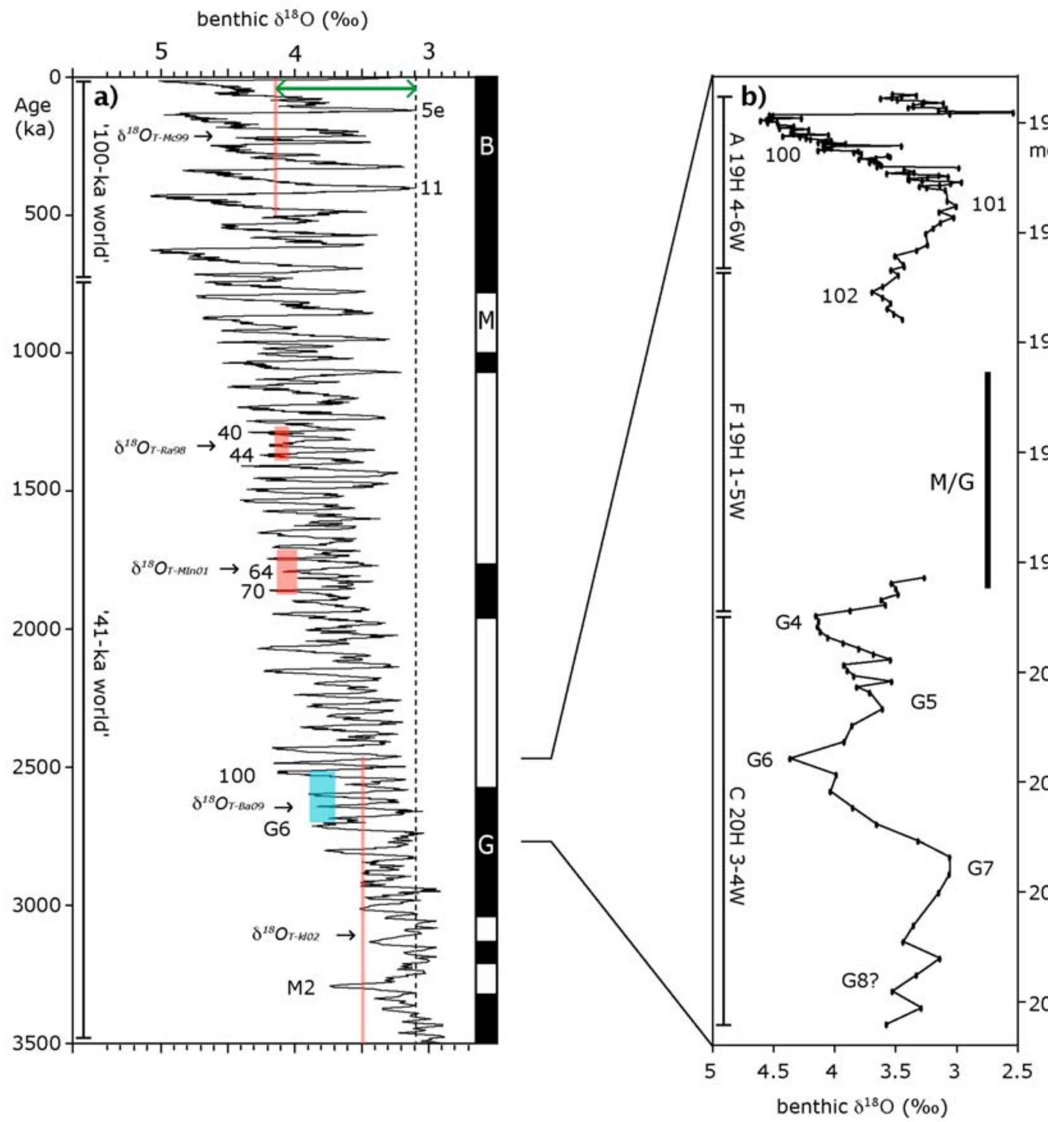

Figure 4. Data from deep-sea sediment cores. (a) LR04 global benthic δ^{18}O stack (equilibrium calcite) for the last 3500 ka [*Lisiecki and Raymo*, 2005]. Vertical pink bars highlight different threshold benthic δ^{18}O values hypothesized for the onset and cessation of North Atlantic ice rafting events during the Plio-Pleistocene ($\delta^{18}O_T$). Labeled arrows indicate $\delta^{18}O_T$ according to *McManus et al.* [1999], $\delta^{18}O_{T\text{-}Mc99}$; *Raymo et al.* [1998], $\delta^{18}O_{T\text{-}Ray98}$; *Mc Intyre et al.* [2001], $\delta^{18}O_{T\text{-}MIn01}$; and *Kleiven et al.* [2002], $\delta^{18}O_{T\text{-}Kl02}$. The vertical blue bar highlights the threshold range in benthic δ^{18}O noted in this study to be associated with the onset and cessation of North Atlantic ice rafting events during MIS G4 and 100 ($\delta^{18}O_{T\text{-}Ba09}$). The green horizontal arrow denotes $\Delta\delta^{18}$O, ~+1.04‰, required to meet $\delta^{18}O_{T\text{-}Mc99}$ during the Last Glacial Cycle [*McManus et al.*, 1999] relative to the Last Interglacial, MIS 5e. The letters "B," "M," and "G" refer to Brunhes, Matuyama, and Gauss paleomagnetic chrons, respectively. (b) Site U1308 benthic δ^{18}O as a function of meters composite depth, mcd. The thick black vertical bar labeled M/G indicates the depth range within which the Brunhes-Gauss paleomagnetic reversal boundary has been identified [*Expedition 303 Scientists*, 2006]. Cores and sections used from holes A, F, and C in the construction of the composite stratigraphy are shown on the left. Numbers denote marine isotope stages referred to in text.

benthic oxygen isotope stratigraphy from DSDP Site 610 used by *Kleiven et al.* [2002] to define the ice volume threshold for region-wide initiation of ice rafting during the late Pliocene ($\delta^{18}O_{T\text{-}Kl02}$) is based on *C. wuellerstorfi* measurements adjusted to equilibrium (δ^{18}O *C. wuellerstorfi* + 0.64‰, K. Kleiven, personal communication, 2007). To help avoid confusion, hereafter, we report all postulated threshold δ^{18}O values adjusted to equilibrium calcite. Thus

$\delta^{18}O_{T\text{-}Mc99}$ becomes +4.14‰ VPDB (3.5‰ + 0.64‰). This simple exercise reveals that, in contrast to conclusions drawn elsewhere, the benthic $\delta^{18}O$ threshold proposed for region-wide Pliocene ice rafting, $\delta^{18}O_{T\text{-}Kl02}$ [*Kleiven et al.*, 2002], is quite different (~0.6‰ VPDB lower) to $\delta^{18}O_{T\text{-}Mc99}$ identified for the last five glacial cycles. We note that some evidence contradicts the findings of *Shackleton and Hall* [1984], suggesting that *Cibicidoides* calcite is, after all, precipitated in isotopic equilibrium with seawater [*Bemis et al.*, 1998]. This possibility does not alter our findings. At issue here is the need for consistency in the way in which data sets are reported and compared rather than the decision on whether or not to adjust individual data sets. Furthermore, while these Pliocene records provide important insights into the initiation of regional ice rafting during intensification of NHG they are not of sufficient resolution to identify suborbital events or to determine their relationship to ice sheet growth. Thus, to understand the suborbital history of subpolar North Atlantic ice rafting and associated surface water variability in the late Pliocene, higher-resolution records are needed.

5. New Records of Pliocene Ice Rafting and Stable Isotopes in Foraminiferal Calcite During MIS G4 and 100, IODP Site U1308

[12] In Figure 5 we present new records of benthic $\delta^{18}O$, $\delta^{13}C$, fragmentation rate of planktic foraminifers, percent foraminifer whole tests and lithic grain counts for MIS 100 and G4 at IODP Site U1308. The concentration of lithics (>150 μm) per gram of bulk dry sediment (lithics g^{-1}) shows large amplitude suborbital time scale variability with frequencies and magnitudes that are comparable between the two glacials (Figures 5h and 5d, respectively). Our $\delta^{13}C$ record from *C. wuellerstorfi* shows the same general pattern as seen in Pleistocene and other late to mid Pliocene records ($\delta^{13}C$ interglacial > glacial) consistent with the changing influence of competing southern- and northern-sourced deep-water masses (Figure 5b). In MIS 100, where the carbon isotope stratigraphy matches the suborbital resolution of our IRD records, cycles in $\delta^{13}C$, likely reflecting oscillations in the mode of meridional overturning circulation, are present even when IRD input is small and when NH ice volume is reduced (see MIS 101, labeled "IG" in Figure 5). During early glacial conditions, the onset of ice rafting events appears to be associated with, and marginally led (over 4 cm, ~550 a) by, an episode of low $\delta^{13}C$, increased fragmentation and decreased concentration of whole tests, reflecting the incursion of more corrosive older southern-sourced waters (compare Figures 5b and 5c to Figure 5d, during "early glacial," EG, MIS 100). A similar relationship has also been reported for MIS 100 for some, but not all, suborbital-scale IRD pulses at ODP Site 981 [*Becker et al.*, 2006], suggesting that a reduction in NADW strength (or southward shift in its locus) and associated decrease in poleward ocean heat transport may contribute to increased ice rafting and not vice versa.

[13] Most importantly for the focus of this paper, the suborbital ice rafting events seen in the glacials that we have studied show threshold behavior with IRD cycles only occurring when benthic $\delta^{18}O$ values exceed about +3.70 VPDB‰ (MIS 100) to +3.90‰ VPDB (MIS G4, ±0.07‰ VPDB at 1σ level; compare Figures 5e and 5a with Figures 5h and 5d). Thus, our high-resolution data sets demonstrate that the band of benthic $\delta^{18}O$ values associated with the onset and cessation of suborbital-scale ice rafting events, at least for glacials MIS G4 and 100 at IODP Site U1308, is less positive than the threshold value identified for the "100 ka (inter)glacial world" at ODP Site 980 ($\delta^{18}O_{T\text{-}Mc99}$ = +4.14‰ VPDB [*McManus et al.*, 1999]).

[14] The offset in values highlighted here for $\delta^{18}O_T$ between the Pliocene and Pleistocene is based on benthic $\delta^{18}O$ data from different sites in the subpolar North Atlantic. However, this offset cannot be accounted for by site-to-site differences in the isotopic composition of deep water ($\delta^{18}O_{sw}$) and temperature. The onset of suborbital ice rafting events in our records for MIS 100 shows a similar association with benthic $\delta^{18}O$ to that observable in published records from DSDP Site 607 (41°00′N) and ODP Site 981 (55°28′N [*Becker et al.*, 2006]). Crucially, ODP Site 981 was drilled on the Feni Drift at almost the same locality and in the same water depth (2127 m) as ODP Site 980 (Figure 3) where the late Pleistocene threshold was defined [*McManus et al.*, 1999]. We therefore conclude that the offset in $\delta^{18}O_T$ that we observe between the late Pliocene and late Pleistocene is a real temporal shift.

6. What Do Different Values Observed for $\delta^{18}O_T$ Represent?

[15] Having established the existence of a real offset in values for $\delta^{18}O_T$ between the late Pleistocene and late Pliocene we next assess the extent to which this finding can be confidently interpreted to reflect changes in the threshold behavior of suborbital ice rafting events in the subpolar North Atlantic.

[16] We must first assess the extent to which differences in $\delta^{18}O_T$ might simply arise from different survivorship histories of drifting icebergs in transit to any given site rather than to the onset and cessation of suborbital iceberg discharge events to the subpolar North Atlantic generally. The flux of IRD to subpolar North Atlantic sediment is primarily dependent on both the production of drifting icebergs, which is likely determined by the response of mature circum–North Atlantic continental ice sheets to millennial climate cycles [*Marshall and Koutnik*, 2006], and the mean position of the polar front. The polar front defines the maximum zone of iceberg melting in the glacial North Atlantic and hence the locus of abundant IRD deposition. Maps of IRD flux to the glacial subpolar North Atlantic indicate that, while the zone of maximum IRD deposition during the LGC occupies a west-southwest trending axis centered on approximately 45 to 50°N, IRD flux increases at all subpolar latitudes north of 40 to 43°N [*Ruddiman*, 1977b, Figure 1]. Hence, while iceberg survivability to a given site might influence the amplitude of suborbital fluctuations in deposition of lithic grains, the onset and cessation of ice rafting cycles in any given subpolar North Atlantic record will track iceberg discharge to the glacial North Atlantic. Evidence in support of this line of reasoning comes from the

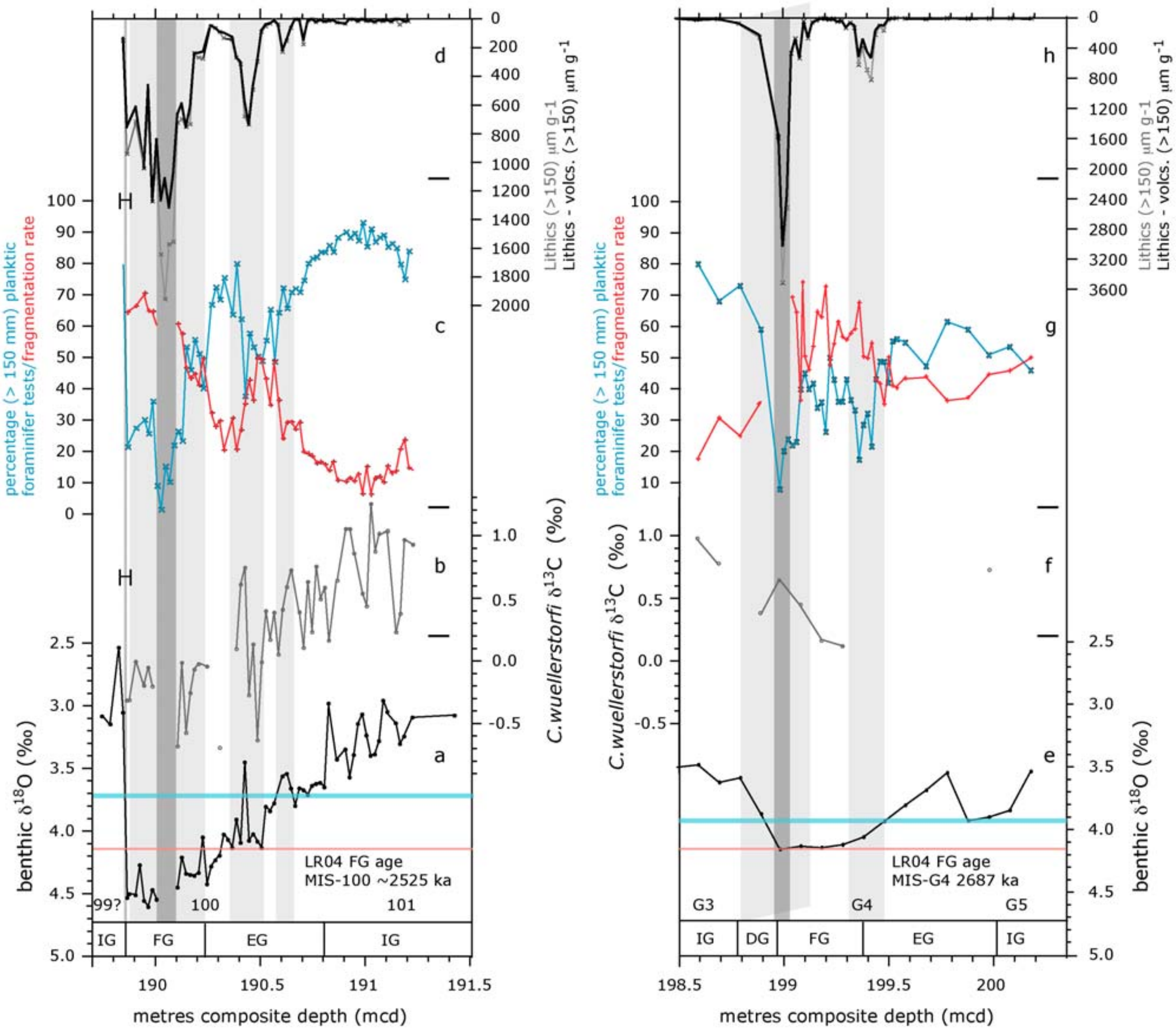

Figure 5. Depth series for (left) MIS 100 and (right) MIS G4 at Site U1308. (a and e) benthic $\delta^{18}O$ (equilibrium calcite); (b and f) *C. wuellerstorfi* $\delta^{13}C$; (c and g) percent, %, foraminifer tests (blue line) and fragmentation rate (red line); and (d and h) lithics (>150 μm) g^{-1} of dry sediment (gray line) and lithics minus volcanic grains (>150 μm) g^{-1} of dry sediment (black line). Horizontal blue bars highlight benthic $\delta^{18}O$ values associated with evidence in lithic counts for the onset of ice rafting events to Site U1308, ~+3.70–3.90‰. Vertical light gray bars pick out ice rafting events during each glacial. Vertical dark gray bars highlight an ash horizon and burrow quartz layer in MIS 100 and MIS G4 sediment, respectively (see main text for details). Bar labeled H highlights stratigraphic hiatus. IG, interglacial; EG, early glacial; FG, full glacial. Data spaced by more than ~1500 ka is not interconnected. We follow standard operating procedure and apply a standard adjustment to equilibrium for $\delta^{18}O$ in *C. wuellerstorfi* (see main text) but refrain from attempting to "correct" $\delta^{13}C$ data generated using the infaunal (albeit shallow infaunal) species *O. umbonatus* to the ambient bottom seawater value and simply report the data from the epifaunal species.

fact that the onset of suborbital ice rafting events in our records for MIS 100 shows a similar association with benthic $\delta^{18}O$ to that observable in published records from DSDP Site 607 (41°00′N) and ODP Site 981 (55°28′N) [*Becker et al.*, 2006]. These two records have been placed onto a common time scale by suborbital tuning of their benthic $\delta^{18}O$ stratigraphies to an independently derived age model [*Becker et al.*, 2006]. Thus, evidence for ice rafting events and their onset during MIS 100 correlate well among sites where data are available and appear to have been broadly coincident over approximately 15° of latitude in the subpolar North Atlantic. This observation is in keeping with the spatial pattern of coarse lithic cycles documented in subpolar North Atlantic sediments of the LGC [*Bond and Lotti*, 1995] and also simulations of NH millennial iceberg delivery to the ocean during this period, which show that the occurrence of suborbital ice rafting events in the subpolar North Atlantic is not governed primarily by SSTs, but by whether the ice sheets are mature enough to advance onto the continental shelf [*Marshall and Koutnik*, 2006].

7. What Does $\delta^{18}O_T$ Mean in Terms of Ice Volume?

[17] $\delta^{18}O_T$ does not appear to change steadily in accordance with the secular increase in benthic $\delta^{18}O$ over the past ~3000 ka (Figure 4). Instead, on the basis of the data now available, we observe two values for $\delta^{18}O_T$ with a shift between approximately 2500 ka and some time during the early Pleistocene (Figure 4a). In principle, the lower value for $\delta^{18}O_T$ that we report for the Pliocene ($\delta^{18}O_{T\text{-}Ba09}$) relative to that reported for the late Pleistocene ($\delta^{18}O_{T\text{-}Mc99}$) can be explained in a number of ways. Four potential explanations are as follows: (1) Suborbital ice rafting events were associated with the same continental ice volume during the two time slices, but there existed warmer North Atlantic bottom waters during the late Pliocene than during the late Pleistocene. (2) Suborbital ice rafting events were associated with the same continental ice volume during the two time slices and there existed a smaller contribution of deep-sea cooling to benthic $\delta^{18}O$ increase during the late Pliocene than during the late Pleistocene. (3) Suborbital ice rafting events were associated with the same continental ice volume during the two time slices and the lower value for $\delta^{18}O_T$ during the Pliocene is attributable to a higher value for the mean isotopic composition of contemporaneous continental ice sheets ($\delta^{18}O_{ice}$) than for the late Pleistocene. (4) Suborbital ice rafting events were associated with a smaller continental ice volume during the late Pliocene than during the late Pleistocene.

7.1. Contribution of Warmer Late Pliocene North Atlantic Deep Waters to Benthic $\delta^{18}O$ for $\delta^{18}O_{T\text{-}Ba09}$

[18] Perhaps the simplest explanation for a Plio-Pleistocene offset in $\delta^{18}O_T$ is that while the onset of suborbital ice rafting events over the past ~3000 ka has been associated with the same (NH) continental ice volume, climate during the late Pliocene was warmer on average relative to the late Pleistocene. In this scenario, isotopic evidence for the same ice volume threshold is offset in Plio-Pleistocene benthic $\delta^{18}O$ only because relatively warmer late Pliocene bottom waters resulted in a lower $\delta^{18}O$ value for $\delta^{18}O_T$ being preserved in benthic foraminiferal calcite. High-latitude North Atlantic records of Plio-Pleistocene surface [*Lawrence et al.*, 2009, Figure 2b] and deep-sea temperature [*Sosdian and Rosenthal*, 2009, Figure 1c] indicate that late Pliocene NH climate following the onset of major NHG during MIS G6 (from ~2750 ka) was warmer on average by ~1–1.5°C. This observation reflects the fact that late Pliocene temperatures during glacial maxima following ~2750 ka are warmer than their late Pleistocene counterparts. However, these data sets also reveal that late Pliocene interglacial temperatures following ~2750 ka are comparable to their late Pleistocene counterparts. The available evidence therefore suggests that only glacial extremes (and not base line conditions prior to the descent into glacials) are offset between the late Pleistocene and late Pliocene [*Lawrence et al.*, 2009]. This conclusion gains some support from emerging boron-isotope-based pCO_2 records [*Seki et al.*, 2010; *Hönisch et al.*, 2009], and also from high-resolution studies of the late Pliocene North Atlantic and Mediterranean [e.g., *Becker et al.*, 2006; *Lourens et al.*, 1996], which suggest that glacial polar waters reached the mid latitudes of the North Atlantic during the late Pliocene (~40°N) and have frequently penetrated into the Mediterranean during cold stages since MIS G6 [*Lourens et al.*, 1996]. Together, these lines of evidence suggest that the offset in $\delta^{18}O_T$ between the late Pleistocene and late Pliocene is unlikely to be explained by warmer North Atlantic bottom waters during the Pliocene.

7.2. Smaller Contribution of Deep-Sea Cooling to Benthic $\delta^{18}O$ for $\delta^{18}O_{T\text{-}Ba09}$

[19] According to the LR04 global benthic stack [*Lisiecki and Raymo*, 2005], amplitude change in benthic oxygen isotopes during the LGC is 1.9‰ VPDB (Figure 4, green horizontal arrow). Over the full glacial cycle, we can assume that ~55% of this signal is attributable to changes in ice volume because it is estimated that, globally, $\delta^{18}O$ of the deep ocean during the Last Glacial Maximum (LGM) was ~1.05‰ higher than today [*Duplessy et al.*, 2002]. Simulations of NH ice sheet growth during the late Pleistocene have demonstrated that the ratio of ice volume to deep-sea temperature contribution to benthic $\delta^{18}O$ records, $\delta^{18}O_R$, is not constant on orbital time scales [*Bintanja et al.*, 2005]. It is also unlikely that $\delta^{18}O_R$ remained fixed at the LGC value for every glacial-interglacial cycle of the past 3000 ka. Yet, it seems unlikely that the evolution of the relative contribution of deep-sea cooling to increases in North Atlantic benthic $\delta^{18}O$ during glacial cycles would have changed systematically in such a way as to generate the offset in $\delta^{18}O_T$ that we observe between the late Pliocene and late Pleistocene. To explain the observed offset in $\delta^{18}O_T$ in this way would require that $\delta^{18}O_R$ to be higher during the late Pliocene relative to the late Pleistocene, in contradiction to independent lines of evidence [*Dwyer et al.*, 1995; *Mudelsee and Ramyo*, 2005; *Sosdian and Rosenthal*, 2009]. Moreover, coupled ice sheet–ocean modeling of the partitioning of ice volume and deep-sea cooling in the LR04 stack over the past 3000 ka, indicates that $\delta^{18}O_R$ may have actually increased over the course of the Plio-Pleistocene and primarily due to an increase in the relative contribution of ice sheet growth to global benthic $\delta^{18}O$ [*Bintanja and van de Wal*, 2008, Figure 1a].

7.3. A Smaller Ice Budget for $\delta^{18}O_{T\text{-}Ba09}$

[20] To test whether the lower value for $\delta^{18}O_T$ that we observe for the late Pliocene is associated with a smaller volume of NH continental ice than that associated with the onset of North Atlantic ice rafting events during the late Pleistocene, we have attempted to ascertain the relationship between $\delta^{18}O_T$ and sea level during these two time slices.

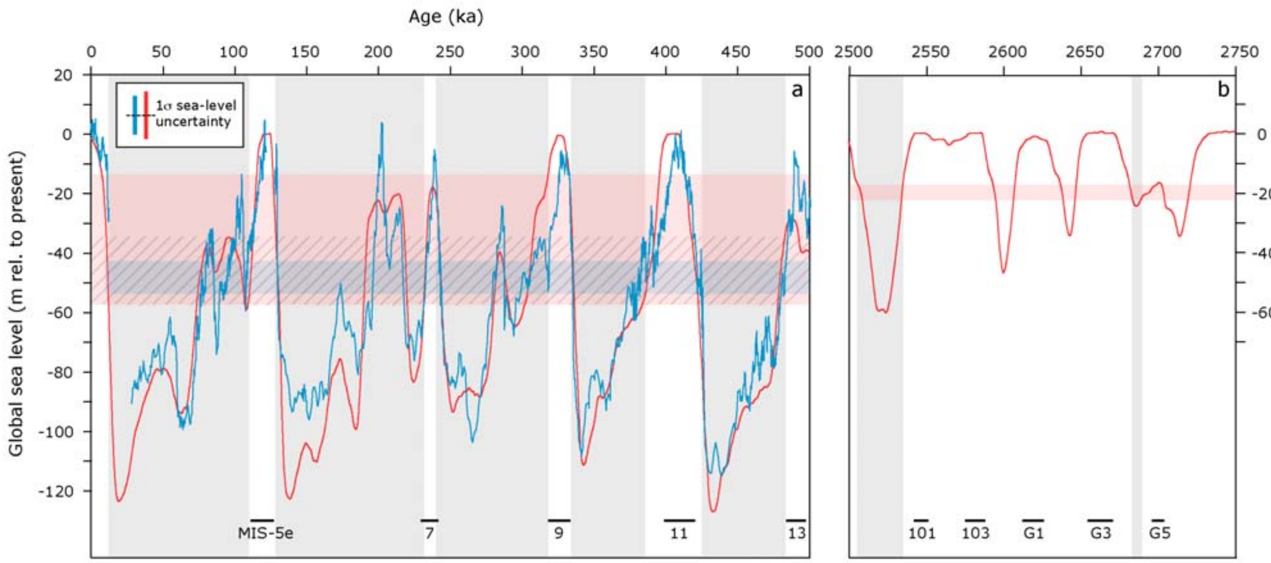

Figure 6. The relationship between the duration of North Atlantic suborbital ice rafting events (vertical gray bars) and global sea level for selected (a) late Pleistocene and (b) late Pliocene (inter)glacials. The red time series is the reconstructed global sea level record of *Bintanja and van de Wal* [2008] for the respective time periods. The blue time series is a three-point moving average of the past ~500 ka of the Red Sea basin forminiferal and bulk sediment δ^{18}O-based global sea level reconstruction (only core GeoTü-KL09) of *Rohling et al.* [2008]. Horizontal red and blue bars denote range of sea level fall for the respective sea level records that is associated with the onset of ice rafting episodes ($\delta^{18}O_T$). Horizontal hatched area in Figure 6a denotes range of sea level fall inferred from the *Bintanja and van de Wal* [2008] record to be associated with $\delta^{18}O_T$ excluding marine isotope stage (MIS) 9 to 8. IRD data for the late Pliocene are taken from this study (MIS G4) and *Becker et al.* [2006] (MIS 100). IRD data for the late Pleistocene are taken from *McManus et al.* [1999]. All data have been placed on the *Bintanja and van de Wal* [2008] sea level chronology via manual graphical correlation of their respective δ^{18}O stratigraphies to the LR04 global benthic stack [*Lisiecki and Raymo*, 2005].

To establish the volume of ice sheet growth (relative to the present) required to meet $\delta^{18}O_T$ during the late Pliocene and late Pleistocene, we have compared the onset of North Atlantic ice rafting events during MIS G4 (from Site U1308, this study), MIS 100 (from Site 981 [*Becker et al.*, 2006]) and the past five glacial cycles (from Site 980 [*McManus et al.*, 1999]) with the Red Sea [*Rohling et al.*, 2008] and *Bintanja and van de Wal* [2008] (BW08) global sea level records.

[21] For our comparison we have tuned both the Red Sea record (core GeoTü-KL09) and the respective North Atlantic stratigraphies to the LR04 age model [*Lisiecki and Raymo*, 2005], which is the time scale used by BW08 (see auxiliary material Figures S1–S3).[1] The results of this exercise are shown in Figure 6. They reveal that $\delta^{18}O_{T\text{-}Mc99}$ is associated with sea level falls ranging from −43 to −53 m (relative to the present) when compared to the Red Sea record (horizontal blue bar in Figure 6a [*Rohling et al.*, 2008]) and −19 to −58 m (relative to the present) when compared to BW08 (horizontal red bar in Figure 7a [*Bintanja and van de Wal*, 2008]. As demonstrated by *Bintanja et al.* [2005] and our comparison here, the inverse modeling approach used in *Bintanja et al.* [2005] and *Bintanja and van de Wal* [2008] is successful at reproducing many features of sea level variability during the late Pleistocene. However, it is noteworthy that there exists a disparity between the Red Sea record and BW08 during the MIS 9–MIS 8 transition. The success of our comparison is partially dependent on the fidelity of the age models used. However, we attribute the majority of this disparity to the explicit exclusion of millennial changes in the *Bintanja et al.* [2005] and *Bintanja and van de Wal* [2008] simulations. Certainly, it is interesting to note that if MIS 9–MIS 8 is excluded from the $\delta^{18}O_T$-BW08 comparison, $\delta^{18}O_{T\text{-}Mc99}$ is associated with a significantly narrower range of −36 to −58 m (relative to the present) that makes for a considerably improved agreement with that determined by comparison of $\delta^{18}O_T$ with *Rohling et al.* [2008].

[22] Given the age model uncertainties embedded in our comparison (see supplementary information) as well as the significant uncertainties in the respective sea level records that we have used (±8 m [*Bintanja and van de Wal*, 2008], ±6.5 m [*Rohling et al.*, 2008] at the 1σ level), it is not currently possible to say with confidence that the benthic δ^{18}O threshold identified by *McManus et al.* [1999] is a specific ice volume threshold. Nevertheless, our work strongly

[1] Auxiliary materials are available in the HTML. doi:10.1029/2009PA001736.

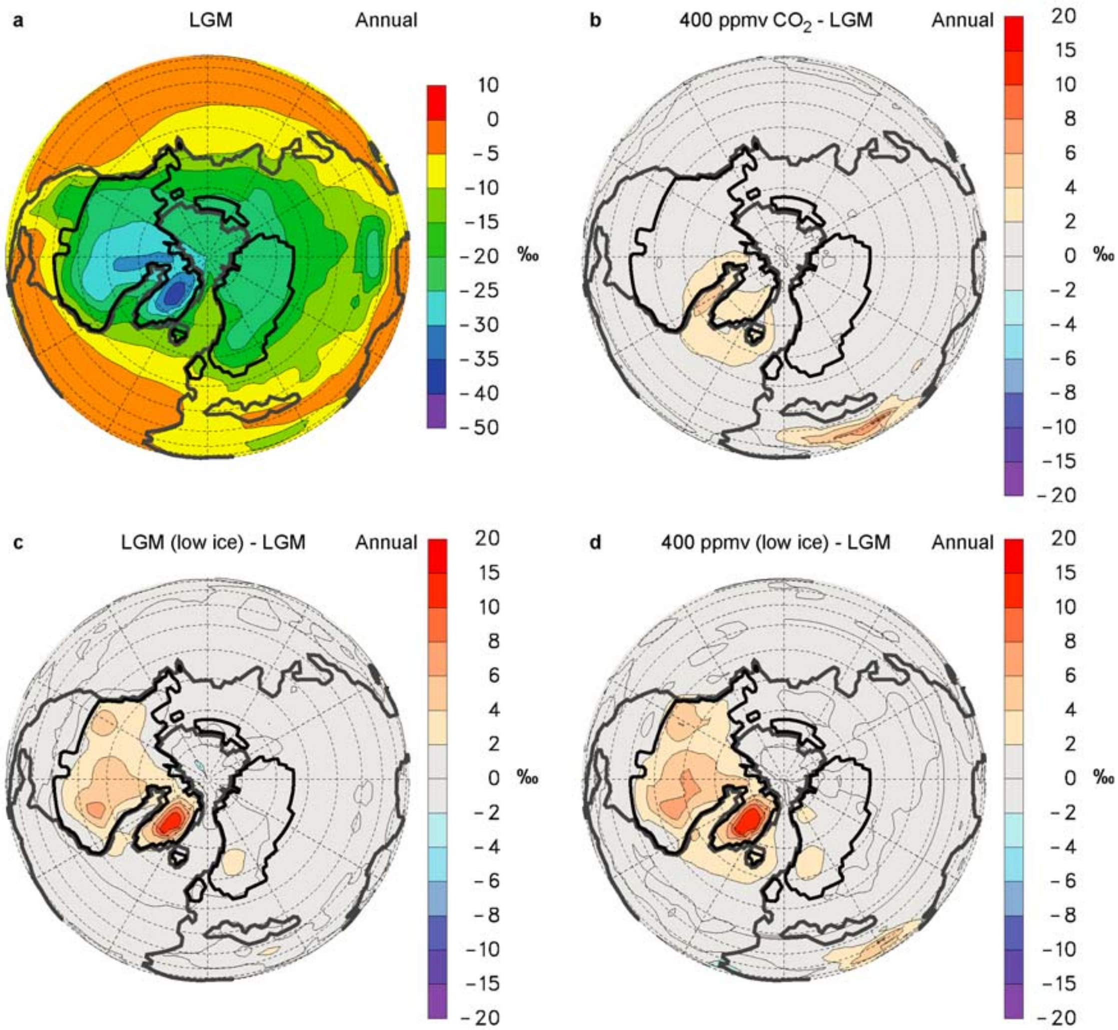

Figure 7. (a) Coupled GCM simulations of annual δ^{18}O precipitation during the Last Glacial Maximum (LGM). The impact of (b) higher pCO$_2$, (c) lower northern hemisphere ice sheets, and (d) higher pCO$_2$ plus lower ice sheets. Model results are averages from the last 10 years of equilibrated 30 year simulations. Gray areas (isotopic differences less than 2‰ VSMOW) are not reported because they are mostly below the level of statistical significance (95% confidence level). LGM ice sheets are shown by the thick black outline.

suggests that the onset of ice rafting events on orbital time scales during the late Pleistocene is associated with a narrow range of sea level fall (~−43 to −53 m below present). It is also apparent that the range of sea level fall associated with meeting $\delta^{18}O_T$ during the late Pliocene glacials MIS G4 and 100 appears to be significantly lower than that associated with meeting $\delta^{18}O_{T\text{-Mc99}}$ (−22 to −17 m, relative to the present, respectively; Figure 6b). We emphasize that because the chronologies assigned here to the benthic δ^{18}O stratigraphies for Sites U1308 and 981 are based on linear sedimentation rates between orbital resolution tie points, age model uncertainty is greater for the late Pliocene IRD records than their late Pleistocene counterparts. Sedimentation rates typically increase during the descent into glacial conditions [*Raymo et al.*, 1998]. Thus, it is possible that the sea level fall associated with meeting $\delta^{18}O_T$ during these two late Pliocene glacials may be a little greater than our linear correlations suggest. Nevertheless, it is unlikely that the uncertainty introduced by the orbital-scale age models used here can account for the entire apparent late Pleistocene–late Pliocene $\delta^{18}O_T$–sea level offset shown in Figures 6a and 6b because this would require an unrealistic increase in sedimentation rate.

[23] The high rate of sea level fall (~−0.5 cm a^{-1}) during the descent into MIS 100 [*Bintanja and van de Wal*, 2008] makes the chronological issues discussed above an important source of uncertainty for our ability to determine the relationship between $\delta^{18}O_T$ and global sea level for the late Pliocene. To reduce this uncertainty, we have further compared the onset of ice rafting at Site 981 during MIS

100 with a new reconstruction of global sea level change for MIS 101 to 95 simulated using the method of *Bintanja et al.* [2005], but based on the benthic δ^{18}O record of ODP Site 967 in the eastern Mediterranean [*Lourens et al.*, 2010]. The amplitude of orbital-scale change in the Site 967 benthic δ^{18}O record (e.g., ~1.1‰ across MIS 100) for this time slice (~2560–2400 ka) is highly comparable to that exhibited by time-equivalent benthic stable oxygen isotope records from the subpolar North Atlantic (i.e., ~1.4 and ~1.2‰, respectively, at Sites 981 and 607 [*Becker et al.*, 2006; *Lourens et al.*, 2010]). The stratigraphy at Site 967 has also been dated independently of phase lags with respect to obliquity tuning as used in LR04, by the correlation of its Ti/Al record with the precessional component of 65°N summer insolation variability, assuming no phase lag [*Lourens et al.*, 2010] and its time scale for MIS 100 exported to the North Atlantic at Sites 981 and 607 [*Becker et al.*, 2006]. Using the tighter chronological control afforded by these suborbital correlations, we find that the onset of ice rafting events during MIS 100 appears to be associated with a sea level fall of only −23 m (relative to the present). These new simulations do not encompass MIS G4, but in this case chronological error is less important to our conclusions because the rate of sea level fall during this glacial appears to have been rather low ($\ll -0.2$ cm a^{-1}) [*Bintanja and van de Wal*, 2008]. Moreover, the maximum sea level fall in BW08 for MIS G4, ~−25 ± 8 m (relative to the present), provides us with an upper limit of the sea level fall associated with the onset of ice rafting events during this glacial. This is significantly smaller than that constrained here for the onset of ice rafting during the late Pleistocene (−43 to −53 m relative to the present, Figure 6a).

7.4. Higher Mean $\delta^{18}O_{ice}$ for $\delta^{18}O_{T\text{-}Ba09}$

[24] The stratigraphic correlations presented in section 7.3 lend support to the view that the Plio-Pleistocene offset in $\delta^{18}O$ that is reported here is traceable to a smaller (NH) ice volume being associated with $\delta^{18}O_{T\text{-}Ba09}$ relative to $\delta^{18}O_{T\text{-}Mc99}$. A corollary of this finding is that smaller ice volumes appear to have been required to initiate ice rafting (ice sheet surging) to the North Atlantic ocean during the late Pliocene than the late Pleistocene. This finding is consistent with the suggestion that NH ice sheets during the late Pliocene and early Pleistocene were lower slung and flowed more readily than their late Pleistocene counterparts as a result of their growth on soft substrates following a long precursor period of sedimentation in a nonglacial weathering regime [*Clark and Pollard*, 1998; *Clark et al.*, 2006; *Hodell et al.*, 2008].

[25] In response to lower-slung ice sheets [*Clark and Pollard*, 1998], and modest latitudinal temperature gradients [*Haywood and Valdes*, 2004] in the late Pliocene relative to the late Pleistocene we might expect Rayleigh distillation processes to give rise to a less extreme value for $\delta^{18}O_{ice}$ growing during Pliocene glacials [cf. *Clark et al.*, 2006; *DeConto et al.*, 2008]. If these mechanisms are capable of producing a Plio-Pleistocene shift in $\delta^{18}O_{ice}$ then a portion of the offset between $\delta^{18}O_{T\text{-}Ba09}$ and $\delta^{18}O_{T\text{-}Mc99}$ may be attributable to changes in latitudinal temperature gradients and the hydrological cycle. To explore this possibility we performed a series of LGM simulations using an isotope-capable GCM.

Table 1. Model Simulations and Relevant Inputs

Experiment	Name	CO_2 Mixing Ratio	Ice Sheet Geometry[a]
1	LGM	200	high-ice
2	high pCO_2	400	high-ice
3	LGM low-ice	200	low-ice
4	high pCO_2 low-ice	400	low-ice

[a]"Low-ice" refers to 50% reduction of ICE-4G LGM [*Peltier*, 1994] Northern Hemisphere ice sheet geometries (high-ice).

[26] The model is based on the current (2008) version of the GENESIS V.3.0 GCM [*Thompson and Pollard*, 1997], with an updated and improved radiation scheme [*Kiehl et al.*, 1998] and embedded water isotopes [*Mathieu et al.*, 2002; *DeConto et al.*, 2008]. The horizontal resolution of the atmospheric component is T31 (~3.75° × ~3.75°) with 18 vertical layers. Surface models including a 50 m slab ocean, dynamic-thermodynamic sea ice, and multilayer models of snow, soil and vegetation are on a finer 2° × 2° grid. The water isotope component [*Mathieu et al.*, 2002] passively tracks the hydrological cycle in the GCM atmosphere and applies relevant fractionation physics during phase transitions. The model considers $^{1}H_2^{18}O$ and $^{1}HD^{18}O$ and accounts for evaporative, condensational, and post condensational processes. Reservoir effects are differentiated over ocean, land (vegetation), and ice sheets. In our preindustrial (control) simulation (supplementary information), the isotopic composition of the ocean surface is prescribed according to modern observations. In the LGM experiments listed below, the ocean is given a uniform global δ^{18}O value of 1.0‰ (VSMOW). The control simulation yields seasonal and spatial distributions of δ^{18}O of precipitation close to present-day observed values, except in the highest elevations of the Antarctic interior, where, as in most GCMs, model surface temperatures are warmer than observed and the δ^{18}O of precipitation is 5–10‰ too heavy [*Mathieu et al.*, 2002; *DeConto et al.*, 2008].

[27] The newly improved version of the isotope-capable GCM used here (GENESIS V. 3.0) includes several modifications for simulating realistic snow mass balances [*Pollard and Thompson*, 1997; *Thompson and Pollard*, 1997] and its performance over ice sheets has been studied extensively [*DeConto et al.*, 2007; *Pollard and Thompson*, 1994; *Pollard and Thompson*, 1997; *Thompson and Pollard*, 1997]. Simulated present-day sea ice distributions and snow mass balances over Antarctica and Greenland have been shown to be among the most realistic of paleoclimatic GCMs [*Pollard*, 2000], and several versions of the model have been used in simulations of Laurentide [*Pollard and Thompson*, 1997] and Antarctic ice sheet variability [*DeConto et al.*, 2007; *DeConto and Pollard*, 2003a; *DeConto and Pollard*, 2003b; *Pollard and DeConto*, 2005].

[28] A series of four experiments (Table 1) were run with varying Northern Hemispheric ice sheet configurations and

atmospheric pCO_2 levels. The ice sheet configurations are based on ICE-4G [*Peltier*, 1994] reconstructions of maximum LGM ice extent (high-ice cases). In two alternative simulations (low-ice cases), ice surface elevations are reduced by 50% over all Northern Hemispheric ice sheets including Greenland. Antarctic ice sheet elevations remain unchanged in all simulations. A relatively cold (LGM) Northern Hemispheric summer orbit is prescribed, with an eccentricity of 0.018994, an obliquity of 22.949°, and climatic precession (the prograde angle between the vernal equinox and perihelion) at 114.52° [*Berger*, 1991]. All simulations were run for 30 model years and are fully equilibrated.

[29] Figure 7 shows the increasing impact of (1) higher pCO_2 (Figure 7b), (2) lower-slung ice sheets (Figure 7c), and (3) higher pCO_2 plus lower-slung ice sheets (Figure 7d) on the isotopic composition of annual precipitation ($\delta^{18}O_p$) falling on NH ice sheets during the LGM. We focus on changes in the δ^{18}O of precipitation over the annual cycle, because they are thought to be representative of the isotopic composition of ice forming during glacials [*Mix and Ruddiman*, 1984]. We focus also on relative changes in $\delta^{18}O_p$ over North American and Eurasian ice sheets because volumetrically, these ice sheets have the greatest influence on ocean chemistry [*Huybrechts*, 2002; *Sima et al.*, 2006]). These results (Figure 7) suggest that changes in ice sheet elevation have a much larger impact on the isotopic composition of accumulating ice than do changes in latitudinal temperature gradients caused by the range of CO_2 concentrations postulated for the Plio-Pleistocene [*Raymo et al.*, 1996; *Sima et al.*, 2006; *Masson-Delmotte et al.*, 2008; *Seki et al.*, 2010; *Hönisch et al.*, 2009]. As shown in Figure 7b, an increase of pCO_2 from 200 ppmv to 400 ppmv has little effect on the isotopic composition of continental $\delta^{18}O_p$, except along the margins of the Laborador Sea/Baffin Bay, where sea ice extents, and hence atmospheric temperatures, are affected most. In contrast, the 50% reduction in ice sheet elevations in the low-ice case (Figure 7c) has a much greater and more widespread effect, increasing $\delta^{18}O_p$ by ~15‰ VSMOW over central Greenland, and ~8‰ VSMOW over the higher elevations of the central Laurentide. However, most of the precipitation falling on the North American and Eurasian ice sheets shows considerably less change (~2–6‰ VSMOW). The additive effect of higher pCO_2 and low-slung ice sheets (Figure 7d) enriches the isotopic composition of annual precipitation slightly, resulting in a more evenly distributed enrichment of $\delta^{18}O_p$ falling on the North American and Eurasian ice sheets, with large changes of between 2 and 6‰ VSMOW evident. Similar spatial patterns of isotopic enrichment are also seen in boreal summer and winter (supplementary information), although the magnitude of the changes is seasonally dependent (compare Figure 7 to Figures S9 and S10). LGM atmospheric general circulation modeling of precipitation over Greenland indicates that annual accumulation over NH ice sheets during full glacial conditions may be dominated by the summer period because the winters are too cold to allow substantial amounts of accumulation [*Werner et al.*, 2001]. This observation does not alter our conclusions because relative changes in $\delta^{18}O_p$ between summer and mean annual differences in our simulations are similar (compare Figure 7 to Figure S9).

8. Did Pliocene Ice Sheets Surge More Readily Than Their Pleistocene Counterparts?

[30] The isotopic differences in $\delta^{18}O_p$ reported in section 7.4 between the model runs for our "high pCO_2 low-ice" and "LGM" cases (typically ~2 to 6‰ VSMOW) (Figure 7d) lend support to the view that the offset between $\delta^{18}O_{T\text{-}Ba09}$ and $\delta^{18}O_{T\text{-}Mc99}$ is partially traceable to comparatively low Pliocene ice sheet elevations. Yet, on the basis of simple mass balance calculations (see supplementary information) we estimate that the effect is much smaller than required (~−7 to −10‰ VSMOW) to account for the entire Plio-Pleistocene offset in $\delta^{18}O_T$. This lends further support to the notion that $\delta^{18}O_{T\text{-}Ba09}$ is associated with a smaller ice volume than $\delta^{18}O_{T\text{-}Mc99}$. Hence, if $\delta^{18}O_T$ primarily reflects a critical threshold in ice volume for the occurrence of North Atlantic suborbital ice rafting events then we conclude that the ice volume threshold proposed for the late Pleistocene [*McManus et al.*, 1999] does not apply to the late Pliocene. In any case, our findings are consistent with the suggestion that NH ice sheets during the late Pliocene and early Pleistocene were lower slung and flowed more readily than their late Pleistocene counterparts. Significantly, perhaps, the best geological evidence in support of the existence of low-slung slippery ice sheets mostly comes from the Pliocene in the form of till deposits (from Missouri, as far south as ~39°N) dated as greater in age than ~1700 ka [*Clark and Pollard*, 1998; *Roy et al.*, 2004; *Balco et al.*, 2005; *Clark et al.*, 2006, Figure 15a; *Raymo and Huybers*, 2008].

9. Conclusions

[31] We present new high-resolution records of benthic δ^{18}O, δ^{13}C, percent foraminifer whole tests, fragmentation rate for planktic foraminifer and lithic grain counts for MIS 100 and G4 at IODP Site U1308. Our lithic grain counts reveal large amplitude suborbital variability in ice rafting to Site U1308 during late Pliocene intensification of NHG. Our records indicate that the relationship between the onset and cessation of late Pliocene North Atlantic suborbital ice rafting events and benthic δ^{18}O ($\delta^{18}O_{T\text{-}Ba09}$ = ~+3.70 to 3.90‰), a proxy for ice volume, is different to that observed for the late Pleistocene ($\delta^{18}O_{T\text{-}Mc99}$, ~+4.14‰ [*McManus et al.*, 1999]). Previous work suggested that the relationship observed between benthic δ^{18}O and North Atlantic ice rafting for the '100-ka (inter)glacial world' is applicable to the '41-ka-dominated' glacial cycles of the late Pliocene [*Kleiven et al.*, 2002]. We reach the opposite conclusion. Our different findings are traceable in part to confusion arising from different ways in which oxygen isotope data are reported and in part to the new light shed on the problem by our more finely resolved records for the Pliocene.

[32] Assuming that the pattern of North Atlantic ice rafting events during Plio-Pleistocene glacials reflects threshold behavior in the Earth system, the offset that we observe between $\delta^{18}O_{T\text{-}Ba09}$ and $\delta^{18}O_{T\text{-}Mc99}$ is most simply explained

by a difference in the volume of ice associated with the two observed thresholds combined with a large difference in the isotopic compositions (~2–6‰ VSMOW) of the growing ice sheets. A series of numerical simulations indicate substantially heavier $\delta^{18}O_{ice}$ for Pliocene NH ice sheets, mostly in response to their reduced elevation relative to the late Pleistocene. We therefore conclude that the observed offset in $\delta^{18}O_T$ is most likely attributable to the existence of lower-slung Pliocene ice sheets that flowed more readily than their late Pleistocene counterparts, in a world with a smaller continental ice volume composed of less isotopically depleted ice. Our findings are consistent with the suggestion that Pliocene ice sheets flowed more readily [*Clark and Pollard*, 1998, and references therein; *Balco et al.*, 2005], on more slippery soft-bedded substrates.

[33] If Pliocene ice sheets were lower slung than their late Pleistocene counterparts, then our findings strongly suggest that the LGC sea level calibration [*Shackleton and Opdyke*, 1973] is not applicable to late Pliocene. In this context, previous combined Mg/Ca ratio and oxygen isotope based estimates of late Pliocene sea level change [*Dwyer et al.*, 1995; *Sosdian and Rosenthal*, 2009] may misrepresent eustatic sea level variability and underestimate the uncertainty associated with their reconstructions.

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