



Deglacial diatom productivity and surface ocean properties over the Bermuda Rise, northeast Sargasso Sea

Isabelle M. Gil,^{1,2,3} Lloyd D. Keigwin,² and Fatima G. Abrantes¹

Received 29 December 2008; revised 11 September 2009; accepted 28 September 2009; published 12 December 2009.

[1] Diatom assemblages document surface hydrographic changes over the Bermuda Rise. Between 19.2 and 14.5 ka, subtropical diatom species and *Chaetoceros* resting spores dominate the flora, as in North Atlantic productive regions today. From 16.9 to 14.6 ka, brackish and fresh water diatoms are common and their contribution is generally coupled with total diatom abundance. This same interval also contains rare grains of ice-rafted debris. Coupling between those proxies suggests that successive discharge of icebergs might have stimulated productivity during Heinrich event 1 (H1). Iceberg migration to the subtropics likely created an isolated environment involving turbulent mixing, upwelled water, and nutrient-rich meltwater, supporting diatom productivity in an otherwise oligotrophic setting. In addition, the occurrence of mode water eddies likely brought silica-rich waters of Southern Ocean origin to the euphotic zone. The persistence of lower-salinity surface water beyond the last ice rafting suggests continued injection of fresh water by cold-core rings and advection around the subtropical gyre. These results indicate that opal productivity may have biased estimates of meridional overturning based on ²³¹Pa/²³⁰Th ratios in Bermuda Rise sediments during H1.

Citation: Gil, I. M., L. D. Keigwin, and F. G. Abrantes (2009), Deglacial diatom productivity and surface ocean properties over the Bermuda Rise, northeast Sargasso Sea, *Paleoceanography*, 24, PA4101, doi:10.1029/2008PA001729.

1. Introduction

[2] Diatoms are common wherever there is water, light and nutrients. Thus, where they are preserved, diatoms are a useful proxy to trace past changes of surface waters. Diatom paleorecords are available from North Atlantic upwelling regions and oceanic margins, but are sparse from the intertropical zone [Abrantes, 2000, 2001; Stabell, 1986]. The paleoclimatic signal there is mainly inferred from *Ethmodiscus* oozes or fresh water diatoms that blew off the African continent [Pokras, 1991; Abrantes, 2003]. Recently, a ~290 cm thick diatom-rich interval representing part of the last deglaciation was reported from the Bermuda Rise [Keigwin and Boyle, 2008] (Figure 1). Here, we describe its characteristics and the possible relationship between the diatoms, local surface ocean freshening, ice rafting and implications for deep-ocean circulation.

2. Oceanographic Setting

[3] The northeast Bermuda Rise is a region of high sediment accumulation that is linked to the Gulf Stream system and deep recirculating gyres [Laine and Hollister, 1981]. Due to the low-nutrient content of surface waters, the primary productivity is low and most opal is remineralized in the upper water column [Heath, 1974]. Sediment trap time series near Bermuda show a constant low opal flux,

with 5.5% and 3.1% opal content of biogenic matter at 3964 m depth and at the surface sediment, respectively [Honjo *et al.*, 1982]. In consequence, few diatoms are found on the seafloor [Maynard, 1976]. The productivity resulting from short-time mesoscale events such as eddies accounts for most of the annual primary productivity in the oligotrophic Sargasso Sea [Michaels, 2007]. Two mesoscale processes enhance productivity today and could have been more important in the past: (1) mode water eddies, known to provide silica and nutrients to the euphotic zone [Krause *et al.*, 2009] and to produce unexpected and major diatom blooms in the Sargasso Sea [McGillicuddy *et al.*, 2007], and (2) cold-core rings, transporting nutrient rich waters and biota from the subpolar gyre [The Ring Group, 1981].

3. Material and Methods

[4] We studied core OCE326GGC6 (GGC6) (33°41.443'N; 57°34.559'W; 4541 m depth) (Figure 1), which is highly correlated by magnetic susceptibility with well-dated companion core OCE326GGC5 [McManus *et al.*, 2004] (GGC5, 33°41.522'N; 57°34.527'W; 4550 m depth) (Figure 2a). The ¹⁴C ages of GGC5 were transferred to GGC6 by aligning the magnetic susceptibility records (Figure 2a). In addition to diatom counts at GGC6, we also report foraminiferal data (from GGC5 and KNR31GPC5 (GPC5), 33°41.2'N; 57°36.9'W; 4583 m depth) and ice-rafted debris (IRD) counts (GGC5 and GGC6). Considering the coordinates of GGC6, GGC5 and GPC5, we assume that they were collected from the same site.

[5] Core GGC6 is 450 cm long and was sampled for diatom analysis every 16 cm from 0 to 150 cm and every 4 cm to the bottom (448 cm). The samples were treated following Abrantes *et al.* [2005]. Diatom assemblages were

¹Marine Geology Department, LNEG, Alfragide, Portugal.

²Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, USA.

³CIMAR Associate Laboratory, Porto, Portugal.

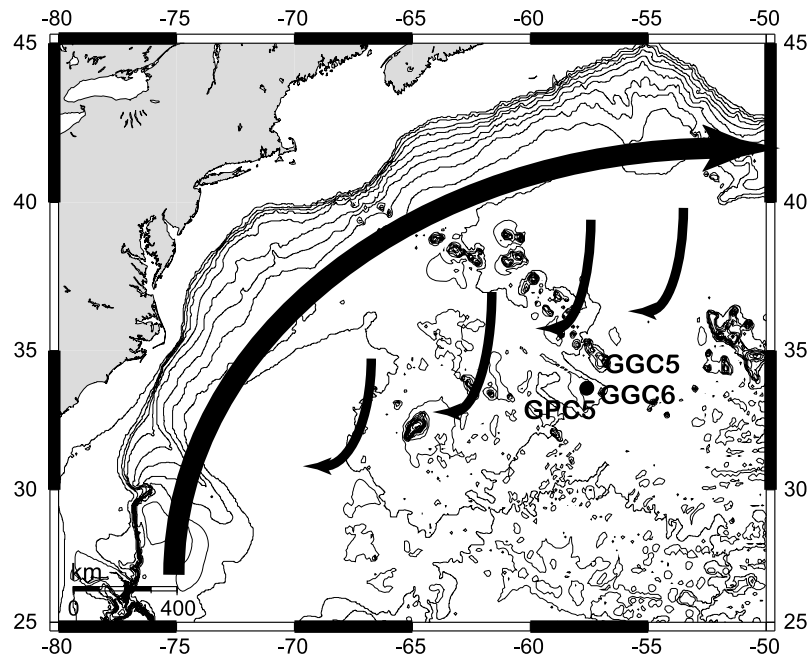


Figure 1. Map showing the location of cores GGC6, GGC5, and GPC5. The large black arrow indicates the pathway of the Gulf Stream, and the other arrows indicate the possible paths for cold-core rings, south of the Gulf Stream.

based on the identification of at least 300 specimens per sample (*Chaetoceros* resting spores are counted, but not included in the total of 300). Considering a counting of 300 specimens, a contribution abundance of 2% of a diatom species/group to the total assemblage is considered a significant contribution, with a confidence level of 95.4% [Galehouse, 1971]. Considering only the species with a minimum contribution of 2% is also justified in transfer function analysis: 300 specimens of planktonic foraminifera are traditionally counted and species of less than 2% abundance are considered too rare for the analysis [Imbrie and Kipp, 1971]. Diatoms were identified only below 168 cm because of low abundances above that level.

[6] Planktonic foraminifera and lithic grains were counted and identified from samples that were wet and dried weighed and further wet sieved to keep the $>63\ \mu\text{m}$ fraction.

[7] The planktonic foraminifera *Globorotalia scitula* and *Neogloboquadrina pachyderma* sinistral (a polar species) were counted at core GPC5 from splits of the $150\ \mu\text{m}$ fraction and presented as percent of the planktonic fauna. These data are from an unpublished census of splits of the planktonic fauna by L. D. Keigwin (unpublished data, 1984).

[8] The lithic grains we found are mostly angular quartz from the $>150\ \mu\text{m}$ fraction. Similar grains are found during Heinrich Events (HE) of isotope stage 3 on the Bermuda Rise and are interpreted as IRD [Keigwin and Boyle, 1999]. IRD counts on core GGC5 from 16.2 to 17.88 ka, including detrital carbonate [Benetti, 2006], were extended with additional counts from GGC6 between 15.2 and 15.5 ka

(200 to 212 cm) and from GGC5 between 14.8 and 15.6 ka (166 to 188 cm).

4. Results

[9] Diatom groups are based on the temperature and salinity indications reported for each diatom species in the literature. The groups are formed in a way to distinguish the diatoms that only and strictly thrive in a determined environment (fresh water, brackish, warm and cold water diatoms groups) from the diatoms species, which have the ability to adapt to environmental changes: for example, the “warm to temperate diatom group” does not include the diatom species that are only found in warm waters. *Chaetoceros* (Ehrenberg) is one of the largest genera of marine planktonic diatoms [Hasle *et al.*, 1996] and comprises numerous species. Abundance and diatom data (species and group) will be available in the PANGAEA database.

[10] Diatoms are common from 19.2 ka (base of the core) until ~ 14.5 ka, when they reached present-day minimum values, and they are most abundant between 16.8 and 15.1 ka (Figure 2b). Most species identified were assembled into groups (see auxiliary material).¹ As expected for the subtropics, warm water diatom species dominate (12%–32%). At 17.7 ka, the contribution of warm diatoms increased relative to warm to temperate species (Figure 3d). *Chaetoceros* resting spores (CRS) are a major contributor to the assemblages (8%–27%) (Figure 3c). The general trend of their contribution is a decrease over the

¹Auxiliary materials are available in the HTML. doi:10.1029/2008PA001729.

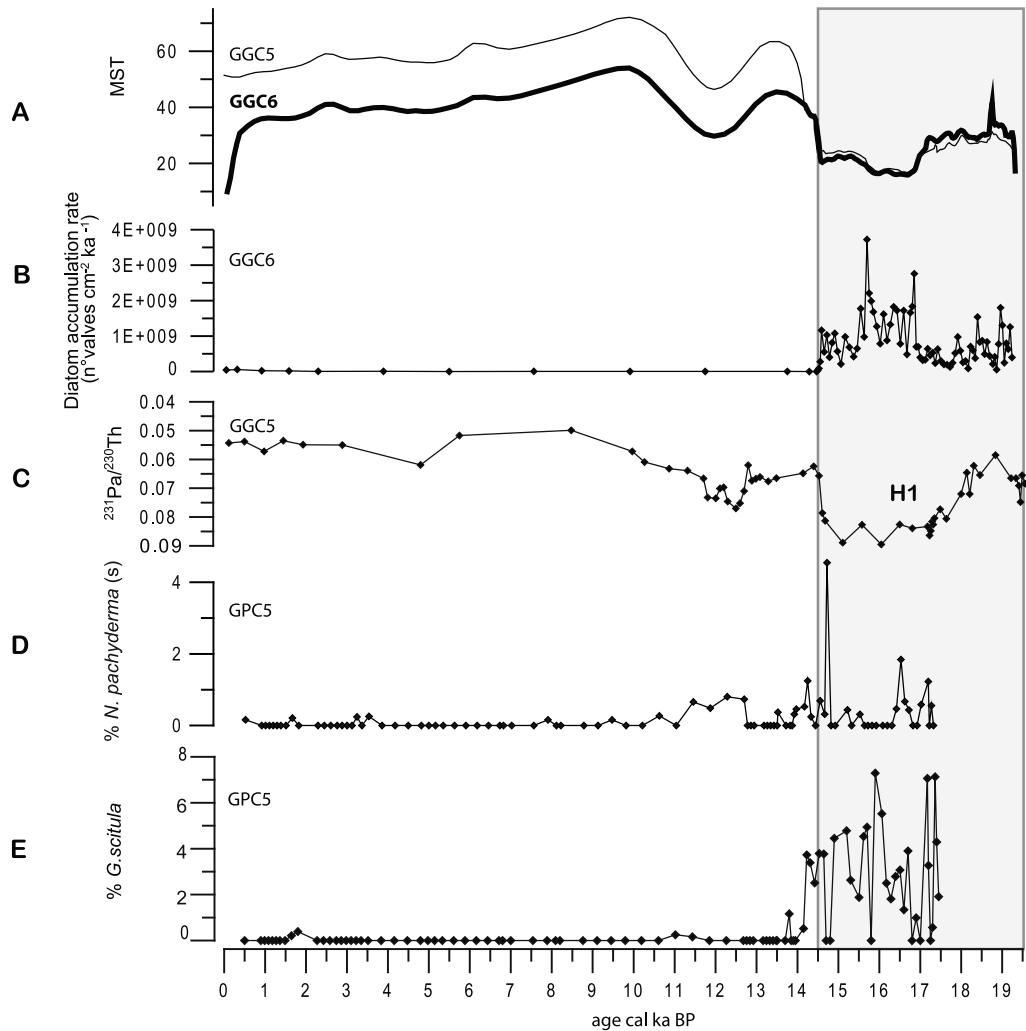


Figure 2. Diverse paleoproxy data from cores GGC6, GGC5, and GPC5. The gray bar highlights the study period detailed in Figure 3. (a) Magnetic susceptibility of cores GGC6 (thick line) and GGC5 (thin line) versus age (ka B.P.); (b) diatom accumulation rate (number of valves $\text{cm}^{-2} \text{ka}^{-1}$) in GGC6; (c) $^{231}\text{Pa}/^{230}\text{Th}$ in GGC5 [McManus *et al.*, 2004]; (d) percentage (%) of *N. pachyderma* (s) in GPC5. (e) Percentage (%) of *G. scitula* in GPC5.

duration of the record, although there are three maxima ~ 15.7 , 16.6 and 18 ka.

[11] Within the CRS group, the contribution of cold water species gradually increases to 15.7 ka and then decreases (Figure 3a). From 16.9 to 14.6 ka (except ~ 16 ka) brackish water diatoms reached significant levels (2.1% – 5.1%) (Figure 3b). Fresh water diatoms (comprising species such as *Nitzschia acicularis* (Kützing) Smith and *Staurosira construens* (Ehrenberg)) reach significant levels at ~ 15.5 , 16.1 and 16.6 ka ($\sim 2.4\%$ each time). Benthic and coastal diatoms reach significant numbers at ~ 18.4 ka (2.7% and 2.3% , respectively) (Figure 3b). Benthic diatoms reach again a significant level (2.3%) at ~ 17.5 ka (Figure 3b).

[12] Quartz grains interpreted to be IRD are found in GGC6 at 15.4 ka and in GGC5 at 15 and 15.9 ka (Figure 3b). The polar planktonic foraminifera *Neogloboquadrina pachyderma* (s) in KNR31GPC5 decreased in abundance suddenly ~ 14 ka, but reappeared

in lower abundance during later cold episodes. In contrast, *Globorotalia scitula* is rarely present after 14 ka (Figure 2).

5. Discussion

[13] Diatoms are especially abundant from 16.8 to 15.1 ka, during the “Mystery Interval” [Denton *et al.*, 2006]. This interval starts with Heinrich Event 1 (H1) (starting ~ 16.8 ka [Bond *et al.*, 2001]) and ended at the onset of the Bølling/Allerød. Its name reflects the contemporary occurrence of features of climate variability that seem incompatible with one another, such as the retreat of European glaciers at a time when other records provide evidence for extreme cold [Denton *et al.*, 2006]. In fact, well-preserved and abundant diatoms in Sargasso Sea sediment add to the mystery, as it does not happen nowadays. However, increased diatom production is also recorded at the same site during Heinrich events 2 and 3 [Lippold *et al.*, 2009]. The occasional

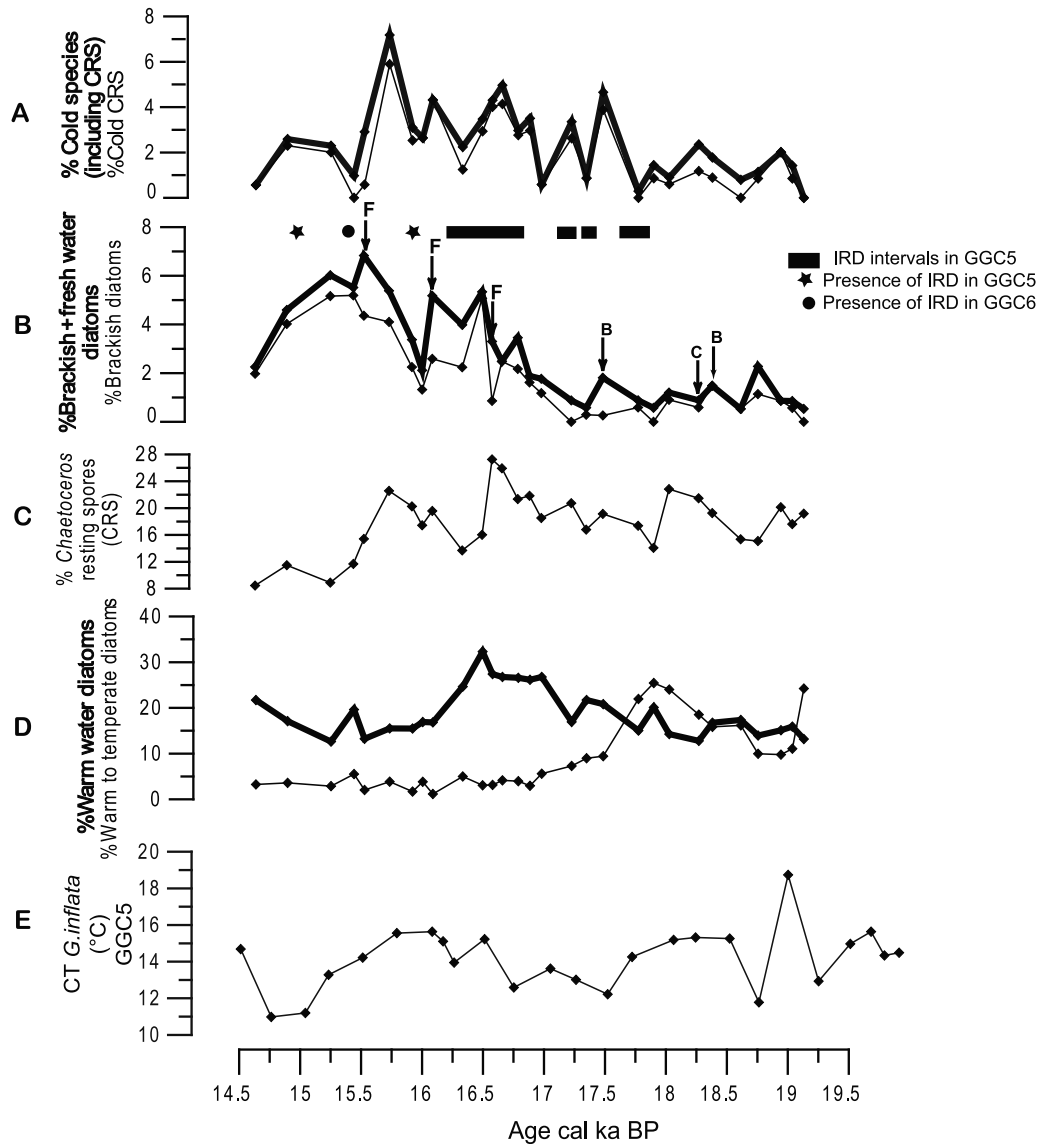


Figure 3. Diatom- and foraminiferal-derived temperature and salinity indicators from GGC6 and GGC5. (a) Contribution of cold water diatom species (thick line) and contribution of cold water *Chaetoceros* resting spores (CRS) only (thin line). (b) Contribution of brackish and fresh water diatoms all together (thick line) and contribution brackish diatom contribution only (thin line) with the letters F, B, and C indicating the significant contributions (above 2%) of fresh water, benthic, and costal diatom species, respectively, in GGC6. The horizontal bold segments indicate IRD intervals in GGC5 [Benetti, 2006], the black stars indicate additional levels of IRD occurring after 16.2 ka B.P. in GGC5, and the plain circle indicates one IRD level in GGC6. (c) Contribution of *Chaetoceros* resting spores in GGC6. (d) Contribution of warm (thick line) and warm to temperate (thin line) diatom species in GGC6. (e) Calcification temperature (CT) in °C derived from Mg/Ca measured on *G. inflata* in GGC5 [Carlson et al., 2008].

presence of the benthic foraminifera *Uvigerina peregrina* (reported also by Boyle and Keigwin [1987] and known from North Atlantic sites with high Total Organic Carbon (TOC) [Miller and Lohmann, 1982]) and observed in >150 μm fraction while searching for IRD) in GGC6 agrees with a relatively high organic carbon rain known from ODP Site 1063 [Keigwin et al., 2001]. The high opal content during H1 corresponds to lower magnetic susceptibility due

to the consequent high porosity and water content of the sediment, as described previously [Mayer, 1982].

5.1. Diatoms and SST

[14] The most common diatoms found are principally CRS and subtropical species (such as *Alveus marinus* (Grunow) Kaczmarek et Fryxell). At ~17.7 ka, the warm water diatom group overtakes the “warm to temperate”

group, which would indicate a relative warming of the surface water (Figure 3d), but the first appearance of IRD is also accompanied by an increase in cold water diatoms (17.7 to 15.5 ka) (Figure 3a), mainly CRS species (*C. furcellatus* Bailey and *C. debilis* Cleve). Considering that the increase of the cold species is sudden (from 0.2% to 4.6%) and the contribution of warm water diatoms remains roughly constant, the decrease of the “warm to temperate” group would express two important aspects of the oceanographic situation: (1) an important temperature contrast between the water masses and (2) a low mixing of the different water masses.

[15] The diatom record differs from the temperature reconstruction based on Mg/Ca in the planktonic foraminifera *Globorotalia inflata* from GGC5 [Carlson et al., 2008] (Figure 3e). Despite the lower resolution of the temperature record based on Mg/Ca, it is possible to observe important misfits in comparison with the diatom assemblage records. From the base of the record to 17.5 ka, the reconstructed temperature values register a large variability, while the records of the warm and cold water diatoms do not. Between ~17.5 ka and ~16.5 ka, when the Mg/Ca SSTs are lower, the contribution of the warm water diatoms tends to increase. Regarding the cold water diatoms, their contribution is negligible (only relatively higher at the beginning and at end of this period) when compared to the contribution of warm water diatoms. Finally, from ~16.5 to 15 ka, the Mg/Ca SSTs are relatively higher, when precisely the contribution of the warm water diatoms decreases and the one of the cold water diatoms increases. The misfits between the foraminifera and diatom records are likely explained by the difference in water depth of their respective habitat. *Globorotalia inflata* is a species known to live between 100 and 400 m depth in the Sargasso Sea during late winter [Anand et al., 2003], whereas diatoms thrive in the euphotic zone (upper ~50 m depth). This means that diatoms reflect the surface ocean (including the environment near icebergs), while the geochemical data likely reflect general conditions in the deeper ice-free Sargasso Sea below where diatoms live. Thus, we would not expect both records to agree.

5.2. Diatoms, Ice, and SSS

[16] Beginning ~16.9 ka, the brackish and fresh water diatoms became significantly represented in Bermuda Rise sediments and IRD mark the first significant discharge of icebergs characterizing H1 in GGC5 [Benetti, 2006] (Figure 3b). HE added an exceptional amount of fresh water into the North Atlantic, but this is thought to have affected principally the subpolar gyre. In fact, IRD in Bermuda Rise sediments during the HE of isotope stage 3 are known to be similarly abundant (~1g⁻¹) [Keigwin and Boyle, 1999] as in core GGC5 [Benetti, 2006]. The evidence for lower salinity and IRD points to successive discharges of icebergs that reached the subtropics and significantly changed the surface salinities. Interestingly, a linear regression between total diatom abundance and fresh and brackish diatom species all together present a positive correlation, with $r = 0.66$ (significance level of $p < 0.0001$). This indicates that these species are an important part of the flora. Their

importance lies not in their low abundance, but in the observation that they are present at all. Although occasional benthic diatoms must have been transported by ice to the Sargasso Sea (benthic diatoms are observed on mineral particles of drifting icebergs in the Southern Ocean (SO) [Smith et al., 2007]), the significance of fresh and brackish water diatoms is less certain. Fossil tests could have been transported, but cold-core rings bearing melting icebergs might have had a sufficiently fresh surface layer to support living populations of these exotic species. Indeed, local positive impacts of melting drifting icebergs on the nutrient budget and on the productivity of the surface water have been observed in the SO [Schwarz and Schodlok, 2009; Smith et al., 2007]. There are no studies dealing with diatom productivity and drifting icebergs from the North Atlantic nowadays, but the SO offers several studies corroborating our scenario. Considering the location in the Sargasso Sea, there is no other way to get substantial and local fresh water input.

5.3. Surface Ocean Scenarios

[17] Cold-core rings in the North Atlantic constitute nowadays a contrasting environment in the Sargasso Sea. They spread south of the Gulf Stream and result from the pinch off of its meanders [The Ring Group, 1981]. Sea surface temperatures (SST) are from 10°C to 16°C at the cold-core ring center whereas these temperatures are usually found as much as 600 m deeper in the Sargasso Sea [The Ring Group, 1981]. Sea Surface Salinity (SSS) of cold-core rings is about 1 per mil lower than the ambient surface water [The Ring Group, 1981].

[18] Cold-core rings present a local productivity increase of 50% (the rings covering 10%–15% of the Sargasso Sea) [The Ring Group, 1981]. Furthermore, diatom patches are found in upwelled cold core of the Gulf Stream [Yoder et al., 1981] and diatoms prevail at the ring margins [Lochte and Pfannkuche, 1987]. Cold-core rings may also carry a characteristic foraminiferal assemblage, including *G. scitula* [Beckmann et al., 1987], which is nowadays absent in plankton tows [Bé, 1960], sediment traps [Anand et al., 2003] and surface sediment [Ruddiman, 1969] from this part of the North Atlantic. The last significant occurrence of this species and *N. pachyderma* (s) on the Bermuda Rise (Figure 2) coincides with the onset of Bølling warming at ~14 ka. Whether fresh water diatoms grew in situ or were transported by icebergs, they must be linked to cold-core rings and to iceberg melting. The variety in diatom species might reflect the structure of the cold-core rings [The Ring Group, 1981]: cold water surrounded by Gulf Stream waters in the Sargasso Sea. These strong SST and SSS contrasts enable a subpolar microcosm to survive transport across the subtropics [The Ring Group, 1981]. The oceanographic processes involved during the “Mystery Interval” were likely amplified by the presence of icebergs, which also causes turbulent mixing [Schwarz and Schodlok, 2009] and likely added an important amount of fresh water. Sancetta [1992] already hypothesized increased diatom productivity during HE as a result of input of dissolved silica and nutrients, but diatom productivity is low within the area of maximum iceberg melting (except above 50°N) during

H1 [Nave *et al.*, 2007]. Our present results are not contradictory as the Bermuda Rise is outside this zone.

[19] The abundant diatom record preserved in the sediment and the enhanced diatom productivity may also be amplified by silica leakage from the SO during deglaciation [Bradtmiller *et al.*, 2007]. Antarctic Intermediate Water (AAIW) could have been brought to the euphotic through mode water eddies. They are associated with a cold surface anomaly and are constituted by a thick lens of subtropical mode water (STMW) (the so-called 18°C water) and lie between a doming seasonal thermocline and a depressed main thermocline [Richardson, 1993; McGillicuddy *et al.*, 1999]. They are suspected to form precisely near the STMW and most of them can even affect water depth as great as 700 m [Richardson, 1993]. The STMW results from the winter cooling of the northern Sargasso Sea surface water and its mean flow is nowadays at 300 m [Worthington, 1959], but descends much deeper when the seasonal stratification sets [Krause *et al.*, 2009].

[20] Assuming the cooling over the North Atlantic and the stratification over the Bermuda Rise during H1 are due to the presence and melting of the icebergs, the STMW was likely flowing deeper. In consequence, mode water eddies affected a deeper part of the water column and likely even interacted with the AAIW, which might have had a shallower mean depth flow by this time. Indeed, the silica rich AAIW likely flooded the deep North Atlantic [Rickaby and Elderfield, 2005; Pahnke *et al.*, 2008] thanks to a sluggish thermohaline circulation [Gherardi *et al.*, 2009] and an increase in production of AAIW [Pahnke *et al.*, 2008; Anderson *et al.*, 2009]. Considering also that winter deep convective mixing (promoting STMW formation) may also reduce the primary production by diminishing the delivery of nutrients to the subsurface [Palter *et al.*, 2005], it appears that a possible explanation for the extraordinary diatom production over the Bermuda Rise during H1 relies also on the possible alternation of periods of storminess (producing an homogeneous STMW) and periods of abrupt cooling leading to strong stratification.

[21] One more very distinguishing feature of deglacial changes over the Bermuda Rise is that the maximum contribution of brackish and fresh water diatoms occurred ~15.5 ka, after the most continuous and abundant inputs in IRD (indicated by the black bars in Figure 3b). The low-salinity anomaly revealed by the diatom record and the reduction in IRD abundance (at 15.4 ka in GGC6 and at 15 and 15.9 ka in GGC5) suggests therefore that low salinity lingered over the Bermuda Rise and/or that the icebergs reaching the Sargasso Sea already released and/or contained less IRD. Our results are supported by modeling experiments that show low-salinity anomalies that lingered by hundreds of years in the subtropical North Atlantic following HE [Levine and Bigg, 2008]. Little is known about the fate of the fresh water resulting from the iceberg melting. The gyre circulation itself or the reduced Atlantic Meridional Overturning Circulation (AMOC) might have prevented an efficient recirculation and incorporation of this water into the deep sea [Lehman *et al.*, 1993]. The sudden decrease in low-salinity water indicators and total diatom

abundance at ~14.5 ka likely marks the establishment of the modern oceanographic conditions that no longer support high diatom productivity.

5.4. Diatoms and AMOC Changes

[22] Finally, our data bear on the extent of AMOC suppression during and after H1. As discussed elsewhere [Keigwin and Boyle, 2008], the presence of biogenic silica on the Bermuda Rise may account for elevated $^{231}\text{Pa}/^{230}\text{Th}$ [McManus *et al.*, 2004], as the period of high $^{231}\text{Pa}/^{230}\text{Th}$ in core GGC5 is nearly identical to the interval of high diatom content (Figure 2b). Because the effect of opal scavenging would drive $^{231}\text{Pa}/^{230}\text{Th}$ toward the production ratio [Chase *et al.*, 2002], high $^{231}\text{Pa}/^{230}\text{Th}$ during and after H1 should not be interpreted as a total shutdown of the AMOC. Evidently, there was shallow overturning in the North Atlantic during H1 [Gherardi *et al.*, 2009]. However, in his review of this paper, Dr. R. F. Anderson pointed out that the implications of high $^{231}\text{Pa}/^{230}\text{Th}$ caused by scavenging of ^{231}Pa by opal production over Bermuda Rise can be reconciled with sedimentary evidence of reduced flow of dense northern source water [e.g., Stanford *et al.*, 2006]. If reduced AMOC enhanced the import of Si rich SO waters which reached the euphotic zone in the North Atlantic by mechanisms described above, then high $^{231}\text{Pa}/^{230}\text{Th}$ would still be a consequence of reduced AMOC, but its response would have been less direct than reported by McManus *et al.* [2004].

6. Conclusion

[23] Subtropical gyres are typically oligotrophic environments, and although diatoms are present in surface waters, they are much less abundant than in upwelling regions and they are rarely preserved on the seafloor. The co-occurrence of increased diatom productivity, IRD, and a mixture of diatom species illustrate the anomalous surface ocean conditions over the Bermuda Rise early in the last deglaciation. The increased diatom production during H1 (from 16.8 to 15.1 ka) over the Bermuda Rise must be due to processes similar to modern cold-core rings (amplified by the presence of melting icebergs) and to mode water eddies interacting with SO source waters. Both would be a great source of nutrients to the Sargasso Sea. In fact, melting icebergs generate diatom productivity in the SO nowadays [Smith *et al.*, 2007] and as in the SO, Bermuda Rise samples contain fresh water and brackish diatoms.

[24] Finally, the covariance between the diatom abundance and the maximum in $^{231}\text{Pa}/^{230}\text{Th}$ [McManus *et al.*, 2004] during and immediately after H1 suggests that the role of biogenic opal on the scavenging of ^{231}Pa has been underestimated. Whereas there is good independent evidence for reduced AMOC during H1, there is no strong evidence for a total shutdown of North Atlantic ventilation.

[25] **Acknowledgments.** Support for this research was provided by the Fundação para a Ciência e Tecnologia through the grant BPD/27214/2006 to I. M. Gil. We also thank an anonymous reviewer and Bob Anderson for his valuable comments and suggestions that improved significantly the discussion.

References

- Abrantes, F. (2000), 200 000 yr diatom records from Atlantic upwelling sites reveal maximum productivity during LGM and a shift in phytoplankton community structure at 185 000 yr, *Earth Planet. Sci. Lett.*, 176, 7–16, doi:10.1016/S0012-821X(99)00312-X.
- Abrantes, F. (2001), Assessing the *Ethmodiscus* ooze problem: New perspective from a study of an eastern equatorial Atlantic core, *Deep Sea Res., Part I*, 48, 125–135, doi:10.1016/S0967-0637(00)00041-8.
- Abrantes, F. (2003), A 340,000 year continental climate record from tropical Africa—News from opal phytoliths from the equatorial Atlantic, *Earth Planet. Sci. Lett.*, 209, 165–179, doi:10.1016/S0012-821X(03)00039-6.
- Abrantes, F., I. Gil, C. Lopes, and M. Castro (2005), Quantitative diatom analyses: A faster cleaning procedure, *Deep Sea Res., Part I*, 52, 189–198, doi:10.1016/j.dsr.2004.05.012.
- Anand, P., H. Elderfield, and M. H. Conte (2003), Calibration of Mg/Ca thermometry in planktonic foraminifera from a sediment trap time series, *Paleoceanography*, 18(2), 1050, doi:10.1029/2002PA000846.
- Anderson, R. F., et al. (2009), Wind-driven upwelling in the Southern Ocean and the deglacial rise in atmospheric CO₂, *Science*, 323(5920), 1443–1448, doi:10.1126/science.1167441.
- Bé, A. W. H. (1960), Ecology of recent planktonic foraminifera: Part 2: Bathymetric and seasonal distributions in the Sargasso Sea off Bermuda, *Micropaleontology*, 6(4), 373–392, doi:10.2307/1484218.
- Beckmann, W., A. Auras, and C. Hemleben (1987), Cyclonic cold-core eddy in the eastern North Atlantic. III. Zooplankton, *Mar. Ecol. Prog. Ser.*, 39, 165–173, doi:10.3354/meps039165.
- Benetti, S. (2006), Late Quaternary sedimentary processes along the western North Atlantic margin, Ph.D. thesis, 188 pp., Univ. of Southampton, Southampton, U. K.
- Bond, G., B. Kromer, J. Beer, R. Muscheler, M. N. Evans, W. Showers, S. Hoffmann, R. Lotti-Bond, I. Hajdas, and G. Bonani (2001), Persistent solar influence on North Atlantic climate during the Holocene, *Science*, 294(5549), 2130–2136, doi:10.1126/science.1065680.
- Boyle, E. A., and L. Keigwin (1987), North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperature, *Nature*, 330, 35–40, doi:10.1038/330035a0.
- Bradt Miller, L. I., R. F. Anderson, M. Q. Fleisher, and L. H. Burckle (2007), Opal burial in the equatorial Atlantic Ocean over the last 30 ka: Implications for glacial-interglacial changes in the ocean silicon budget, *Paleoceanography*, 22, PA4216, doi:10.1029/2007PA001443.
- Carlson, A. E., et al. (2008), Subtropical Atlantic salinity variability and Atlantic meridional circulation during the last deglaciation, *Geology*, 36(12), 991–994, doi:10.1130/G25080A.1.
- Chase, Z., et al. (2002), The influence of particle composition and particle flux on scavenging of Th, Pa and Be in the ocean, *Earth Planet. Sci. Lett.*, 204, 215–229, doi:10.1016/S0012-821X(02)00984-6.
- Denton, G. H., W. S. Broecker, and R. B. Alley (2006), The mystery interval 17.5 to 14.5 kyrs ago, *PAGES News*, 14(2), 14–16.
- Galehouse, J. S. (1971), Point counting, in *Procedures in Sedimentary Petrology*, edited by R. E. Carver, pp. 385–407, Wiley-Interscience, New York.
- Gherardi, J.-M., L. Labeyrie, S. Nave, R. Francois, J. F. McManus, and E. Cortijo (2009), Glacial-interglacial circulation changes inferred from ²³¹Pa/²³⁰Th sedimentary record in the North Atlantic region, *Paleoceanography*, 24, PA2204, doi:10.1029/2008PA001696.
- Hasle, G. R., E. E. Syversten, K. A. Steidinger, and K. Tangen (1996), Marine diatoms, in *Identifying Marine Diatoms and Dinoflagellates*, edited by C. R. Tomas, pp. 5–386, Academic, San Diego, Calif.
- Heath, G. R. (1974), Dissolved silica and deep-sea sediments, in *Studies in Paleo-oceanography*, edited by W. W. Hay, *Spec. Publ. Soc. Econ. Paleontol. Mineral.*, 20, 77–93.
- Honjo, S., et al. (1982), Sedimentation of lithogenic particles in the deep ocean, *Mar. Geol.*, 50, 199–220, doi:10.1016/0025-3227(82)90139-6.
- Imbrie, J., and N. G. Kipp (1971), A new micro-paleontological method for quantitative paleoclimatology: Application to a late Pleistocene Caribbean core, in *Late Cenozoic Glacial Ages*, edited by K. K. Turekian, pp. 71–181, Yale Univ. Press, New Haven, Conn.
- Keigwin, L. D., and E. A. Boyle (1999), Surface and deep ocean variability in the northern Sargasso Sea during marine isotope stage 3, *Paleoceanography*, 14(2), 164–170, doi:10.1029/1998PA0900026.
- Keigwin, L. D., and E. A. Boyle (2008), Did North Atlantic overturning halt 17,000 years ago?, *Paleoceanography*, 23, PA1101, doi:10.1029/2007PA001500.
- Keigwin, L. D., et al. (2001), *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 172, Ocean Drill. Program, College Station, Tex.
- Krause, J. W., M. W. Lomas, and D. M. Nelson (2009), Biogenic silica at the Bermuda Atlantic Time-series Study site in the Sargasso Sea: Temporal changes and their inferred controls based on a 15-year record, *Global Biogeochem. Cycles*, 23, GB3004, doi:10.1029/2008GB003236.
- Laine, E. P., and C. Hollister (1981), Geological effects of the Gulf Stream system on the northern Bermuda Rise, *Mar. Geol.*, 39, 277–310, doi:10.1016/0025-3227(81)90076-1.
- Lehman, S. J., et al. (1993), Transport of freshwater into the deep ocean by the conveyor, in *Ice in the Climate System*, edited by W. R. Peltier, pp. 187–209, Kluwer Acad., Norwell, Mass.
- Levine, R. C., and G. R. Bigg (2008), Sensitivity of the glacial ocean to Heinrich events from different iceberg sources, as modeled by a coupled atmosphere-iceberg-ocean model, *Paleoceanography*, 23, PA4213, doi:10.1029/2008PA001613.
- Lippold, J., J. Grützner, D. Winter, Y. Lahaye, A. Mangini, and M. Christl (2009), Does sedimentary ²³¹Pa/²³⁰Th from the Bermuda Rise monitor past Atlantic meridional overturning circulation?, *Geophys. Res. Lett.*, 36, L12601, doi:10.1029/2009GL038068.
- Lochte, K., and O. Pfannkuche (1987), Cyclonic cold-core eddy in the eastern North Atlantic. II. Nutrients, phytoplankton and bacterioplankton, *Mar. Ecol. Prog. Ser.*, 39, 153–164, doi:10.3354/meps039153.
- Mayer, L. A. (1982), Physical properties of sediment recovered on deep sea drilling project Leg 68 with the hydraulic piston corer, *Initial Rep. Deep Sea Drill. Proj.*, 68, 365–382.
- Maynard, N. G. (1976), Relationship between diatoms surface sediments of the Atlantic Ocean and the biological response and physical oceanography of overlying waters, *Paleobiology*, 2, 99–121.
- McGillicuddy, D. J., Jr., R. Johnson, D. A. Siegel, A. F. Michaels, N. R. Bates, and A. H. Knapp (1999), Mesoscale variations of biogeochemical properties in the Sargasso Sea, *J. Geophys. Res.*, 104(C6), 13,381–13,394, doi:10.1029/1999JC900021.
- McGillicuddy, D. J., Jr., et al. (2007), Eddy/wind interactions stimulate extraordinary mid-ocean plankton blooms, *Science*, 316(5827), 1021–1026, doi:10.1126/science.1136256.
- McManus, J. F., R. Francois, J.-M. Gherardi, L. D. Keigwin, and S. Brown-Leger (2004), Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, 428, 834–837, doi:10.1038/nature02494.
- Michaels, A. F. (2007), Highly active eddies, *Science*, 316(5827), 992–993, doi:10.1126/science.1140059.
- Miller, K. G., and G. P. Lohmann (1982), Environmental distribution of recent benthic foraminifera on the northeast United States continental slope, *Geol. Soc. Am. Bull.*, 93(3), 200–206, doi:10.1130/0016-7606(1982)93<200:EDORBF>2.0.CO;2.
- Nave, S., L. Labeyrie, J. Gherardi, N. Caillon, E. Cortijo, C. Kissel, and F. Abrantes (2007), Primary productivity response to Heinrich events in the North Atlantic Ocean and Norwegian Sea, *Paleoceanography*, 22, PA3216, doi:10.1029/2006PA001335.
- Pahnke, K., et al. (2008), Abrupt changes in Antarctic Intermediate Water circulation over the past 25,000 years, *Nat. Geosci.*, 1(12), 870–874, doi:10.1038/ngeo360.
- Palter, J. B., et al. (2005), The effect of advection on the nutrient reservoir in the North Atlantic subtropical gyre, *Nature*, 437, 687–692, doi:10.1038/nature03969.
- Pokras, E. M. (1991), Source areas and transport mechanisms for fresh-water and brackish-water diatoms deposited in pelagic sediments of the equatorial Atlantic, *Quat. Res.*, 35(1), 144–156, doi:10.1016/0033-5894(91)90101-A.
- Richardson, P. L. (1993), A census of eddies observed in North Atlantic SOFAR float data, *Prog. Oceanogr.*, 31(1), 1–50, doi:10.1016/0079-6611(93)90022-6.
- Rickaby, R. E. M., and H. Elderfield (2005), Evidence from the high-latitude North Atlantic for variations in Antarctic Intermediate Water flow during the last deglaciation, *Geochem. Geophys. Geosyst.*, 6, Q05001, doi:10.1029/2004GC000858.
- Ruddiman, W. F. (1969), Recent planktonic foraminifera: Dominance and diversity in North Atlantic surface sediments, *Science*, 164(3884), 1164–1167, doi:10.1126/science.164.3884.1164.
- Sancetta, C. (1992), Primary production in the glacial North Atlantic and North Pacific oceans, *Nature*, 360, 249–251, doi:10.1038/360249a0.
- Schwarz, J. N., and M. P. Schodlok (2009), Impact of drifting icebergs on surface phytoplankton biomass in the Southern Ocean: Ocean colour remote sensing and in situ iceberg tracking, *Deep Sea Res., Part I*, 56, 1727–1741.
- Smith, K. L., et al. (2007), Free-drifting icebergs: Hot spots of chemical and biological enrichment in the Weddell Sea, *Science*, 317(5837), 478–482, doi:10.1126/science.1142834.

- Stabell, B. (1986), Variations of diatom flux in the eastern equatorial Atlantic during the last 400,000 years (meteor cores 13519 and 13521), *Mar. Geol.*, 72, 305–323, doi:10.1016/0025-3227(86)90125-8.
- Stanford, J. D., E. J. Rohling, S. E. Hunter, A. P. Roberts, S. O. Rasmussen, E. Bard, J. McManus, and R. G. Fairbanks (2006), Timing of meltwater pulse 1a and climate responses to meltwater injections, *Paleoceanography*, 21, PA4103, doi:10.1029/2006PA001340.
- The Ring Group (1981), Gulf Stream cold-core rings: Their physics, chemistry, and biology, *Science*, 212(4499), 1091–1100, doi:10.1126/science.212.4499.1091.
- Worthington, L. V. (1959), The 18-degree water in the Sargasso Sea, *Deep Sea Res.*, 5, 297–305.
- Yoder, J. A., L. P. Atkinson, T. N. Lee, H. H. Kim, and C. R. McClain (1981), Role of the Gulf Stream frontal eddies in forming phytoplankton patches on the outer southeastern shelf, *Limnol. Oceanogr.*, 26(6), 1103–1110.
-
- F. G. Abrantes and I. M. Gil, Marine Geology Department, LNEG, Apartado 7586, P-2721-866 Alfragide, Portugal. (igil@whoi.edu)
- L. D. Keigwin, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA.