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Contribution of changes in opal productivity and nutrient distribution in the coastal upwelling systems to Late Pliocene/Early Pleistocene climate cooling

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Abstract. The global Late Pliocene/Early Pleistocene cooling (∼ 3.0–2.0 million years ago – Ma) concurred with extremely high diatom and biogenic opal production in most of the major coastal upwelling regions. This phenomenon was particularly pronounced in the Benguela upwelling sys-

during the Late Pliocene/Early Pleistocene cooling. A pronounced shift of the centres of opal deposition linked to diatom productivity from the North Pacific and Southern Ocean to the regions of the low-latitude eastern boundary currents occurred between 3.0 and 2.4 Ma followed by a second shift from the coastal upwelling regions to the Southern Atlantic Ocean between 2.4 and 2.0 Ma (Cortese et al., 2004), where they are largest in the modern ocean. In the BUS, this extremely high siliceous productivity event, the so-called MDM (Berger et al., 2002; Lange et al., 1999; Pérez et al., 2001), is represented by sedimentary biogenic opal concentrations of up to 60 % along the continental slopes off Namibia and southwest Africa between the Walvis Ridge and Cape Town (Wefer et al., 1998). This high biogenic opal abundance coincided with the first appearance of the diatom flora (e.g. *Thalassiothrix antarctica*) presently prevailing in the Southern Ocean (Berger et al., 2002; Lange et al., 1999; Pérez et al., 2001), which paradoxically evolved under warm surface water conditions and weak upwelling intensity (Etourneau et al., 2009).

The reasons for these changes are particularly intriguing because similar and synchronous maxima in biogenic opal production have also been reported in other coastal upwelling systems, such as off California and Mauritania (Janecek, 2000; Tiedemann, 1991), at which a similarly dramatic SST decline of about 3–4 °C was recorded in all these three areas (Dekens et al., 2007; Etourneau et al., 2009; Herbert and Schuffert, 1998; Liu et al., 2008; Marlow et al., 2000). The most pressing questions regarding the cause of the MDM are therefore: (i) how to explain the co-occurrence of the largest amount of Antarctic diatom mats concomitant with surface water conditions in the coastal upwelling regions characterized by high temperatures and stratification most of the year during the Late Pliocene/Early Pleistocene?(ii) Why did the decline of the Antarctic diatom flora accompany the decrease in opal production when upwelling intensified at the beginning of the early Pleistocene, despite the fact that the development of the upwelling system should have enhanced nutrient supply and sustained higher diatom productivity?

Fig. 1. Silicate and nitrate concentrations (µmol l^{-1}) at 10 m water depth (**A**, **B**) and of water masses along a latitudinal transect (**C**, **D**). White dots indicate the location of the Benguela ODP Sites 1082 and 1084 as well as the South Atlantic Site 1091. As illustrated here, Si is supplied in the BUS by the upwelling of SAMW, subantarctic mode waters, and AAIW, Antarctic Intermediate Waters, formed in the Southern Ocean, in the polar frontal system zone. AC, Angola Current; AGC, Aghulas Current; BC, Benguela Current; BCC, Benguela Coastal Current; BOC,

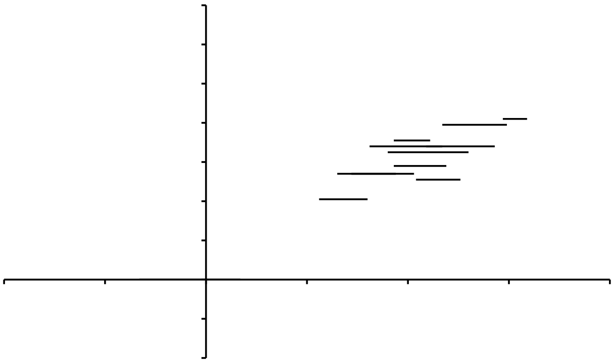

Fig. 2. δ^{30}Si (‰) versus δ^{29}Si (‰).

3 Results and discussion

3.1 Diatom-δ^{30}Si evidence for local changes in Si cycling during the MDM

During the MDM interval, the diatom δ^{30}Si values at Site 1082 were in the range expected for an upwelling area but experienced large amplitude changes ranging between ~0 and 1.7 ‰ (Fig. 3). Except between 3.2 and 3.0 Ma where two high δ^{30}Si values concurred with low BSi MAR, the lowest values around 0.3–0.4 ‰ generally coincide with minima in BSi content (< 20 %) and BSi MAR (1 g cm^{-2} ka^{-1}), whereas the highest values (up to 1.4–1.7 ‰) correspond to the maxima of opal accumulation rates and diatom productivity (respectively, up to 60 % and 4 g cm^{-2} ka^{-1}). Given that diatoms fractionate the dissolved seawater Si by −1.1 ‰ during incorporation into their opal frustules (De La Rocha et al., 1997), δ^{30}Si values of up to 1.7 ‰ suggest that the surface waters in which the diatoms grew had isotopic values in the order of 2.8 ‰, whereas surface water values of only 1.5 ‰ are reconstructed for minimum diatom signatures of 0.4 ‰. For comparison, the bulk sediment and diatom-bound δ^{15}N was at a minimum during the MDM and revealed very light values around 0–2 ‰, while after 2.4 Ma, the values increased to 4–5 ‰ during the Mid-Pleistocene (Fig. 3) (Etourneau et al., 2009; Robinson and Meyers, 2002) which is consistent with the distribution found in the modern situation (Pichevin et al., 2005).

Similar to nitrogen isotopes (δ^{15}N), silicon isotopes (δ^{30}Si) are fractionated during utilisation in the surface waters in a way that the lighter Si isotopes are preferentially incorporated into diatom frustules (De La Rocha et al., 1998, 2000; Brzezinski et al., 2002; Varela et al., 2004), which leaves the ambient seawater enriched in the heavier isotopes. The degree of depletion thus determines the magnitude of isotopic fractionation and together with water mass mixing controls the dissolved signature in ambient seawater (e.g. Reynolds et al., 2006).

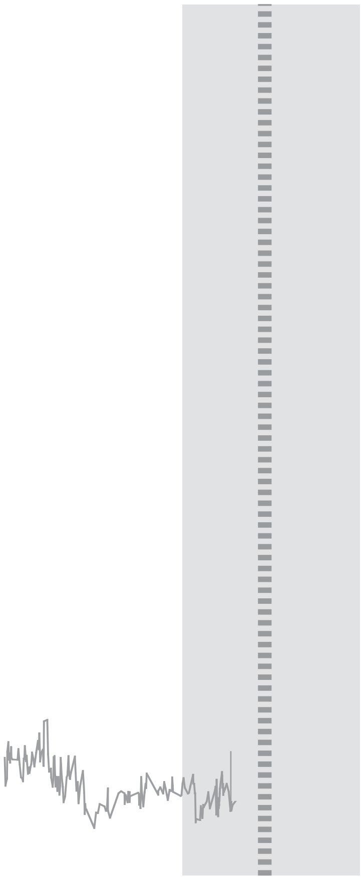

Fig. 3. (A) BSi content (%) at the Benguela Sites 1082 (green)

studies (Berger et al., 2002; Lange et al., 1999; Marlow et al., 2000; Pérez et al., 2001) already suggested that the MDM was characterized by long periods of stratified conditions of surface waters and weak upwelling activity in the BUS due to the presence of *T. antarctica*, a mat-forming diatom species which requires extended periods of stable and stratified conditions for growing and consuming a significant amount of silicate to build their mats. In addition, the appearance of a smaller dominant diatom species (*Chaetoceros* spores), typical of modern upwelling conditions, only occurred when the MDM declined after 2.4–2.0 Ma (Fig. 2). Very few other diatom species (e.g. *Chaetoceros radicans* and *C. cinctus*) that developed under strong upwelling conditions sporadically appeared during the MDM and might have accompanied the excursions to light δ^{30}Si as recorded during episodes of BSi MAR low between 3.0 and 2.0 Ma. However, those species remained globally rare and were more abundant in the south than in the north of the BUS (Lange et al., 1999). The distribution of these diatom species along the Namibian coast implies that developed upwelling cells were restricted to the south of the BUS that corroborates the low presence or absence of *T. antarctica* in the southern BUS (Lange et al., 1999). Warm sea surface temperatures (SST) and low north–south SST gradients reported through the Benguela region between 3.0 and 2.4 Ma (see Fig. 3b) also support our assumption that the upwelling activity was globally weak (Etourneau et al., 2009, 2010; Marlow et al., 2000) during the period of increased diatom and opal productivity. This hypothesis is furthermore consistent with pollen data obtained at Site 1082, demonstrating that wetter conditions dominated the first part of the MDM due to weak upwelling activity and enhanced moisture transport to the continent (Dupont et al., 2005; Dupont, 2006). Thus, compelling evidence from different markers strongly supports weak coastal upwelling systems off Namibia as a feature of the Late Pliocene/Early Pleistocene cooling.

According to this line of evidence, we therefore emphasize that nutrient feeding the biological production in the BUS originated from SAWM and AAIW. The diatom species (*T. antarctica*) dominating the MDM probably mostly migrated through the water column to abundantly feed, in response to global cooling, from shallow thermocline and nutricline waters (Etourneau et al., 2009; Philander and Fedorov, 2003) where the silicate and nitrate-rich SAMW and AAIW circulated. In addition, the growth of diatoms and other phytoplankton species (e.g. coccolithoporids), that were not able to migrate through the water column like diatoms, may have also been sporadically fueled by brief seasonal episodes of mixing supplying nutrients to the surface as well as surface water advection through the Coastal Benguela Current transporting nutrients from the active upwelling cells located south of the BUS.

This weak BUS activity during the MDM was probably closely linked to the system of atmospheric pressure cells that engender strong trade winds and cause the upwelling of deep waters along the Namibian coast. The ocean high pressure cells were likely situated at a more southern position during the MDM owing to warmer atmospheric temperatures and a reduced ice cap over Antarctica, as well as warmer SSTs in the Southern Ocean (Martinez-Garcia et al., 2010; Pollard and Deconto, 2009). This would probably have maintained the South Atlantic high pressure cells far from the African continental low and prevented the formation of strong trade winds along the shore, especially in the northern part of the BUS, thus resulting in a weaker upwelling intensity. The transfer of advected nitrate and silicate-rich waters masses supplying the BUS during the MDM associated with long periods of stratified surface water conditions therefore offered favourable and stable conditions for the growth of mat-forming diatoms, the most significant contributor of biogenic opal production. Silicate was nearly completely consumed by *T. antarctica* for growing and building their extensive mats and progressively became the limiting factor of growth as revealed by their high δ^{30}Si signatures. By comparison, the utilisation of nitrate remained much lower owing to its important supply from the Southern Ocean to the BUS, as indicated by the low δ^{15}N signatures.

In contrast, between 2.4 and 2.0 Ma, the MDM decline is marked by a pronounced reduction of the biogenic opal production and a progressive shift from Antarctic species (*T. antarctica*) to upwelling diatom species (*Chaetoceros* spores and cetae) in concert with the intensification of the upwelling activity (Etourneau et al., 2009; Marlow et al., 2000) and the establishment of a strong meridional atmospheric circulation (Etourneau et al., 2010). In parallel, the polar frontal system in the Southern Ocean developed (Liu et al., 2008), which was accompanied by an increase in Southern Ocean siliceous productivity (Cortese et al., 2004), likely stimulated by increasing iron supply via aeolian dust (Martinez-Garcia et al., 2011). This probably restricted the Si and N export to low-latitude upwelling regions and the nutrient access to upwelling productivity. Concomitantly, the δ^{30}Si values and biogenic opal content in the BUS decreased, whereas the δ^{15}N increased (Fig. 1). Any significant effects of the changing diatom community on Si isotope fractionation are unlikely because no abrupt variation in δ^{30}Si was observed during the shift from Antarctic to upwelling species. We also suppose that the δ^{30}Si was essentially only measured on biogenic opal derived from the same diatom species as those producing the organic matter on which bulk δ^{15}N was measured. Hence, taken together these arguments strongly suggest that the decreasing δ^{30}Si and increasing δ^{15}N values have been the consequence of a diminished utilisation of silica and a higher utilisation of nitrate, respectively, although the δ^{15}N may be also overprinted by heavier isotope signatures circulating through the nutricline (Etourneau et al., 2009). The development of the upwelling system likely triggered a more continuous supply of both bioavailable Si and N. However, the higher utilisation of nitrate likely drove the latter to become a more limiting factor for phytoplankton growth, as illustrated by the higher δ^{15}N values, so that Si utilisation by local diatoms was reduced and led to lighter δ^{30}Si values. In addition, the production of *Chaetoceros* spores instead of *T. antarctica* probably drove to reduced silicate utilisation and biogenic opal production as the former species generates less biogenic opal owing to their smaller size than the Antarctic species and its lesser competitive capacity against the other micro-organisms.

4 Implications for Late Pliocene/Early Pleistocene atmospheric CO_2

The shift of the centres of opal deposition away from the polar oceans and low latitudes most likely had fundamental consequences for the carbon cycle and as such may also have had dramatic consequences for global climate. Although further (modelling) studies are required to precisely quantify the impact of the MDM on global climate, our new results suggest a possible scenario linking this event, nutrient distribution and opal productivity changes in the BUS, as well as in other coastal upwelling systems (Fig. 4), to the Late Pliocene/Early Pleistocene global climate shift (Fig. 5),

Here, we propose that during the warm Pliocene (A) vertical mixing probably dominated the surface waters of the polar oceans (Haug et al., 1999; Sigman et al., 2004), whereas conditions of warm surface waters and weak upwelling activity governed the low-latitude eastern boundary current re-

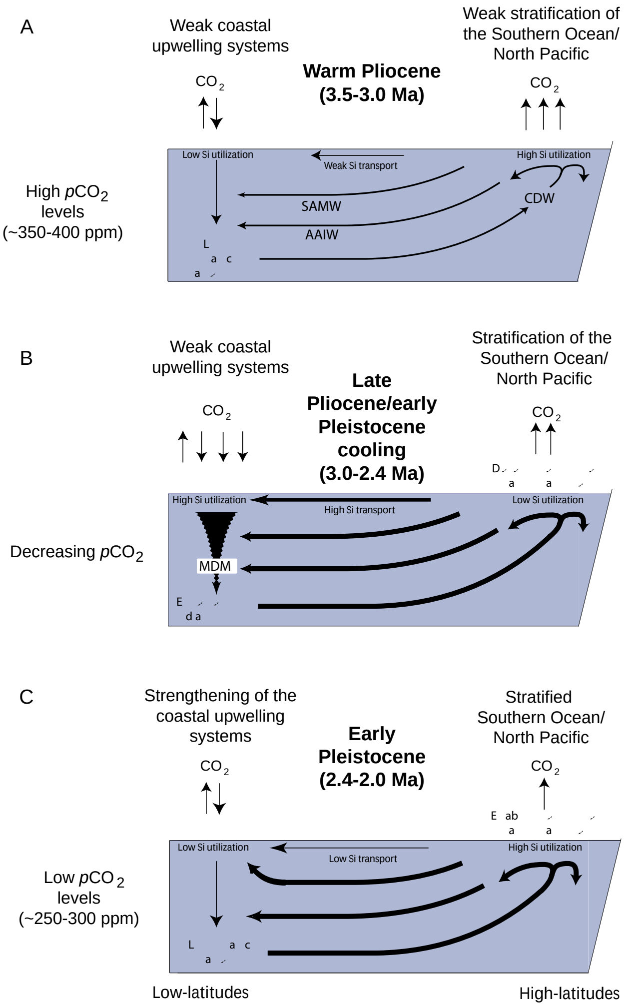

Fig. 5. Proposed scenario representing the Si transport from the Southern Ocean/North Pacific towards upwelling systems and its utilisation by phytoplankton productivity during **(A)** the warm Pliocene, **(B)** the Late Pliocene/Early Pleistocene cooling and **(C)** the early Pleistocene. CDW, Circumpolar Deep Water.

regions and thus promoted high primary productivity rates in these areas. For most of the time, on both decadal scales but probably also during the annual cycle, stratification prevailed within the eastern boundary currents between 2.8 and 2.4 Ma but was interrupted by most likely short seasonal episodes of mixing supplying nutrients, thus fuelling bioproductivity

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