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Hydrographic variations in deep ocean temperature over the mid-Pleistocene transition

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A B S T R A C T

During the mid-Pleistocene transition the dominant 41 ka periodicity of glacial cycles transitioned to a quasi-100 ka periodicity for reasons not yet known. This study investigates the potential role of deep ocean hydrography by examining oxygen isotope ratios in benthic foraminifera. Oxygen isotope records from the Atlantic, Pacific and Indian Ocean basins are separated into their ice volume and local temperature/hydrography components using a piece-wise linear transfer function and a temperature calibration. Although our method has certain limitations, the deep ocean hydrography reconstructions show that glacial deep ocean temperatures approached freezing point as the mid-Pleistocene transition progressed. Further analysis suggests that water mass reorganisation could have been responsible for these temperature changes, leading to such stable conditions in the deep ocean that some obliquity cycles were skipped until precessional forcing triggered deglaciation, creating the apparent quasi-100 ka pattern. This study supports previous work that suggests multiples of obliquity cycles dominate the quasi-100 ka glacial cycles with precession components driving deglaciations.

1. Introduction

The mid-Pleistocene transition (MPT) (700–1250 ka (Clark et al., 2006)), also known as the mid-Pleistocene revolution (MPR) or early-middle Pleistocene transition (EMPT), denotes the transition in the periodicity of glacial cycles from 41 ka to approximately 100 ka (Pisias and Moore, 1981). Several authors have found that the '100 ka' spectral peak, dominant during the last 700 ka, is in fact a broad peak centred around 100 ka rather than a sharp peak (Wunsch, 2003; Clark et al., 2006). Eccentricity has peaks at 96, 125 and 413 ka and therefore does not explain this broad peak centred around 100 ka. A detailed study of the MPT using advanced time-series analysis techniques concluded that the MPT was a multi-staged transition from a more linear to a more non-linear climate state (Mudelsee and Stattegger, 1997). Elucidating the climate system mechanisms that caused the MPT and subsequent quasi-100 ka glacial cycles has been one of the major focal points of recent palaeoclimate research, but a consensus has not yet been reached (e.g. Clark et al., 2006; Siddall et al., 2010a; Ganopolski and Calov, 2011; Huybers, 2011; McClymont et al., 2013).

The quasi-100 ka glacial cycles might result from non-linear interactions between eccentricity, which itself has a ~100 ka pacing (96 ka and 125 ka), and some component within the climate system (e.g. Berger, 1999). This could have been triggered by an increase in ice volume that caused the ice sheets to behave non-linearly in response to eccentricity (Imbrie et al., 1993). For example, Abe-Ouchi et al. (2013) demonstrate that a quasi-100 ka pattern in ice volume is produced when the North American ice sheet is forced by only eccentricity and precession (obliquity kept fixed at 23.5°), when atmospheric CO_2 concentrations are 190–230 ppm and isostatic rebound is included in their model. One explanation for an ice volume increase through the Pleistocene is the regolith hypothesis, whereby exposed bedrock allowed the ice sheets to thicken because of greater frictional forces (Clark and Pollard, 1998). Bintanja and van de Wal (2008) attribute the MPT to the merging of North American ice sheets. Other climate system components could also have interacted non-linearly with eccentricity, such as sea ice (e.g. Gildor and Tziperman, 2000; Tziperman and Gildor, 2003; de Garidel-Thoron et al., 2005), ocean circulation (Berger and Jansen, 1994; Siddall et al., 2010a) or the carbon cycle (e.g. Raymo et al., 1997; Köhler and Bintanja, 2008). Alternatively the quasi-100 ka glacial cycles might not arise from the ~100 ka

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eccentricity cycle despite their coincidental timing, meaning that the quasi-100 ka periodicity of these glacial cycles is a multiple of precession (e.g. Paillard, 1998; Ridgwell et al., 1999; Maslin and Ridgwell, 2005) or obliquity forcing (e.g. Huybers and Wunsch, 2005; Huybers, 2007, 2009; Liu et al., 2008; Siddall et al., 2010a). A further possibility is that the quasi-100 ka glacial cycles resulted from the synchronisation of the 413 ka eccentricity cycle with the climate system (Rial, 2004; Rial et al., 2013). We note that this brief introduction only includes a relatively small selection of hypotheses about the MPT; this is purely for the sake of brevity and we refer the reader to comprehensive reviews such as Clark et al. (2006) and McClymont et al. (2013).

Oxygen isotope ratios in foraminiferal calcium carbonate, from marine sediment cores, have been a highly useful tool for studying glacial cycles because these records are long, continuous and simultaneously document fluctuations in temperature, ice volume and hydrography (Chappell and Shackleton, 1986). These $\delta^{18}O$ data contain a component related to the $\delta^{18}O$ of the seawater in which the foraminifera grew ($\delta^{18}O_{sw}$) (Shackleton, 1967) and one related to their growth temperature ($\delta^{18}O_T$) (Urey, 1947). In turn, $\delta^{18}O_{sw}$ consists of a global component due to fluctuations in continental ice sheet volume ($\delta^{18}O_{ice}$) and a local component resulting from hydrographic variations caused by changes in the properties and positions of deep ocean water masses ($\delta^{18}O_{local}$) (e.g. Waelbroeck et al., 2002). When considering variability on the timescale of glacial cycles, it can be preferable to use a benthic oxygen isotope record ($\delta^{18}O_b$) rather than one from planktic foraminifera, because the deep-sea habitat does not experience such strong regional and seasonal temperature and freshwater balance fluctuations (Lear et al., 2000; Lear, 2007; Sosdian and Rosenthal, 2009). For this reason, $\delta^{18}O_b$ is sometimes approximated to contain only the $\delta^{18}O_{ice}$ and $\delta^{18}O_T$ components (e.g. Siddall et al., 2010a).

The separation of the ice volume component ($\delta^{18}O_{ice}$) from the deep ocean temperature component ($\delta^{18}O_T$) within $\delta^{18}O_b$ can reveal the relative contributions of these two factors through time (Chappell and Shackleton, 1986). The comparison of $\delta^{18}O_b$ records from different locations can be used to study hydrography by comparing the $\delta^{18}O_{sw}$ of different sites (Waelbroeck et al., 2002). Therefore, the analysis of $\delta^{18}O_b$ records that span the MPT and come from a variety of locations could provide clues about changes in ice volume, deep ocean temperature and deep ocean hydrography that occurred during the MPT (Clark et al., 2006).

The temperature component of $\delta^{18}O_b$ can be estimated using an independent temperature recorder such as Mg/Ca ratios in foraminiferal calcium carbonate (e.g. Lear et al., 2000; Billups and Schrag, 2003; Sosdian and Rosenthal, 2009). However, very few data are currently available at adequate resolution and duration for use in isolating the temperature component of $\delta^{18}O_b$ over the MPT (Sosdian and Rosenthal, 2009; Elderfield et al., 2012). Furthermore, one of these records has been called into question (Sosdian and Rosenthal, 2010; Yu and Broecker, 2010). In addition, Mg/Ca data are influenced by carbonate saturation state and basin-dependent calibration (Lear et al., 2002; Elderfield et al., 2006; Marchitto et al., 2007). Therefore alternative estimates of deep ocean temperature from multiple sites with multi-million year duration remain useful and necessary. A recent study by Elderfield et al. (2012) used Mg/Ca ratios from a sediment core on Chatham Rise to separate the effects of decreasing temperature and increasing global ice volume on oxygen isotope ratios. These results were interpreted to suggest that the MPT was initiated by an abrupt increase in Antarctic ice volume at 0.9 Ma. We compare the Mg/Ca temperature record of Elderfield et al. (2012) with our estimates later in this paper.

Here we follow the method of Chappell and Shackleton (1986), Waelbroeck et al. (2002), Cutler et al. (2003) and Siddall et al. (2010a), whereby $\delta^{18}O_{ice}$ is calculated and subtracted from $\delta^{18}O_b$ to yield $\delta^{18}O_{T+local}$ ($\delta^{18}O_{T+local} = \delta^{18}O_T + \delta^{18}O_{local}$), which can then be used to estimate temperature with an empirically-derived equation such as that of Shackleton (1974). Our study combines two approaches, one by Waelbroeck et al. (2002) who considered multiple sites covering the last four climatic cycles, and the other by Siddall et al. (2010a) who selected a $\delta^{18}O_b$ record that spans the mid-Pleistocene transition. In this way, the method facilitates the study of ice volume, deep ocean temperature and deep ocean hydrography variations through the MPT.

2. Material and methods

2.1. $\delta^{18}O_b$ records

We used published $\delta^{18}O_b$ records from marine sediment cores drilled during the Deep Sea Drilling Project (DSDP) and Ocean Drilling Project (ODP), as well as one composite record compiled by Shackleton et al. (1990) that was previously used by Siddall et al. (2010a) for similar purposes. The $\delta^{18}O_b$ records cover the time period of interest, have adequate time resolution during the stadials and interstadials used for the ice volume calibration and are reasonably globally-distributed (Fig. 1). However, sites in the Pacific are mostly limited to the eastern equatorial region and the South China Sea. Fig. 2 displays each of these $\delta^{18}O_b$ records as they appeared in their original publications (i.e. with no species

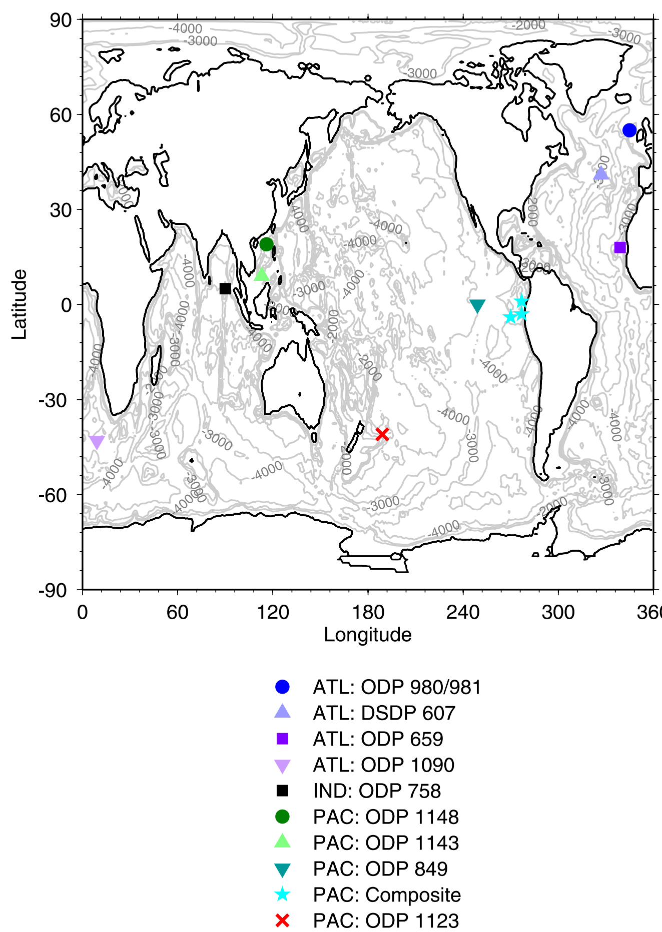

Fig. 1. World map displaying the site locations of the $\delta^{18}O_b$ records used by this study. Colours correspond to Table 1. Latitudes and longitudes are shown in Table 1 along with references for those values. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)