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The role of orbital forcing in the Early Middle Pleistocene Transition

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ABSTRACT

The Early Middle Pleistocene Transition (EMPT) is the term used to describe the prolongation and intensification of glacial–interglacial climate cycles that initiated after 900,000 years ago. During the transition glacial–interglacial cycles shift from lasting 41,000 years to an average of 100,000 years. The structure of these glacial–interglacial cycles shifts from smooth to more abrupt ‘saw-toothed’ like transitions. Despite eccentricity having by far the weakest influence on insolation received at the Earth's surface of any of the orbital parameters; it is often assumed to be the primary driver of the post-EMPT 100,000 years climate cycles because of the similarity in duration. The traditional solution to this is to call for a highly nonlinear response by the global climate system to eccentricity. This ‘eccentricity myth’ is due to an artefact of spectral analysis which means that the last 8 glacial–interglacial average out at about 100,000 years in length despite ranging from 80,000 to 120,000 years. With the realisation that eccentricity is not the major driving force a debate has emerged as to whether precession or obliquity controlled the timing of the most recent glacial–interglacial cycles. Some argue that post-EMPT deglaciations occurred every four or five precessional cycle while others argue it is every second or third obliquity cycle. We review these current theories and suggest that though phase-locking between orbital forcing and global ice volume may occur the chaotic nature of the climate system response means the relationship is not consistent through the last 900,000 years.

1. Introduction

The Early Middle Pleistocene Transition (EMPT; previously known as the Mid-Pleistocene Transition or Revolution; Berger and Jansen, 1994; Head et al., 2008) is the last major ‘event’ or transition in a secular trend towards more intensive global glaciation that characterizes the late Cenozoic (Zachos et al., 2001). The earliest recorded onset of significant regional glaciation during the Cenozoic was the widespread continental glaciation of Antarctica at about 34 Ma (e.g., Zachos et al., 2001; Huber and Nof, 2006; Sijp et al., 2009). Perennial sea ice cover in the Arctic has occurred throughout the past 14 Ma (Darby, 2008; Schepper et al., 2014). Glaciation in the Northern Hemisphere lagged behind, with the earliest recorded glaciation on Greenland occurring before about 6 Ma (e.g., Larsen et al., 1994; Thiede et al., 2011). Schepper et al. (2014) have identified a number of key Pliocene glacial events which may have been global and occurred at 4.9–4.8 Ma, ~4.0 Ma, ~3.6 Ma and ~3.3 Ma. It is not until the Pliocene–Pleistocene transition that the long-term cooling trend culminates in the glaciation of Northern Europe and North America around 2.6 Ma (Maslin et al., 1998). The extent of glaciation did not evolve smoothly after this, but instead was characterized by periodic advances and retreats of ice sheets on a hemispherical scale – the ‘glacial–interglacial cycles’.

The EMPT is the marked prolongation and intensification of glacial–interglacial climate cycles initiated sometime between 900 and 650 ka (Fig. 1). Before the EMPT, global climate conditions appear to have responded primarily to the obliquity orbital periodicity (Imbrie et al., 1992; Tiedemann et al., 1994; Clark et al., 2006; Elderfield et al., 2012) through glacial–interglacial cycles with a mean periodicity of ~41 kys. After about 900 ka, starting with Marine Oxygen Isotope Stage (MOIS) 22, glacial–interglacial cycles start to occur with a longer duration and a marked increase in the amplitude of global ice volume variations (Elderfield et al., 2012; Rohling et al., 2014). The increase in the contrast between warm and cold periods may also be in part due to the extreme warmth of many of the post-EMPT interglacial periods as similar interglacial conditions can only be found at ~1.1 Ma, ~1.3 Ma and before ~2.2 Ma. Fig. 2 shows time-series analysis of the ODP 659 (Tropical East Atlantic ocean) benthic foraminifera oxygen isotope record spanning the EMPT (Mudelsee and Stattegger, 1997). The analysis suggests the EMPT was a two-step process with the first transition at

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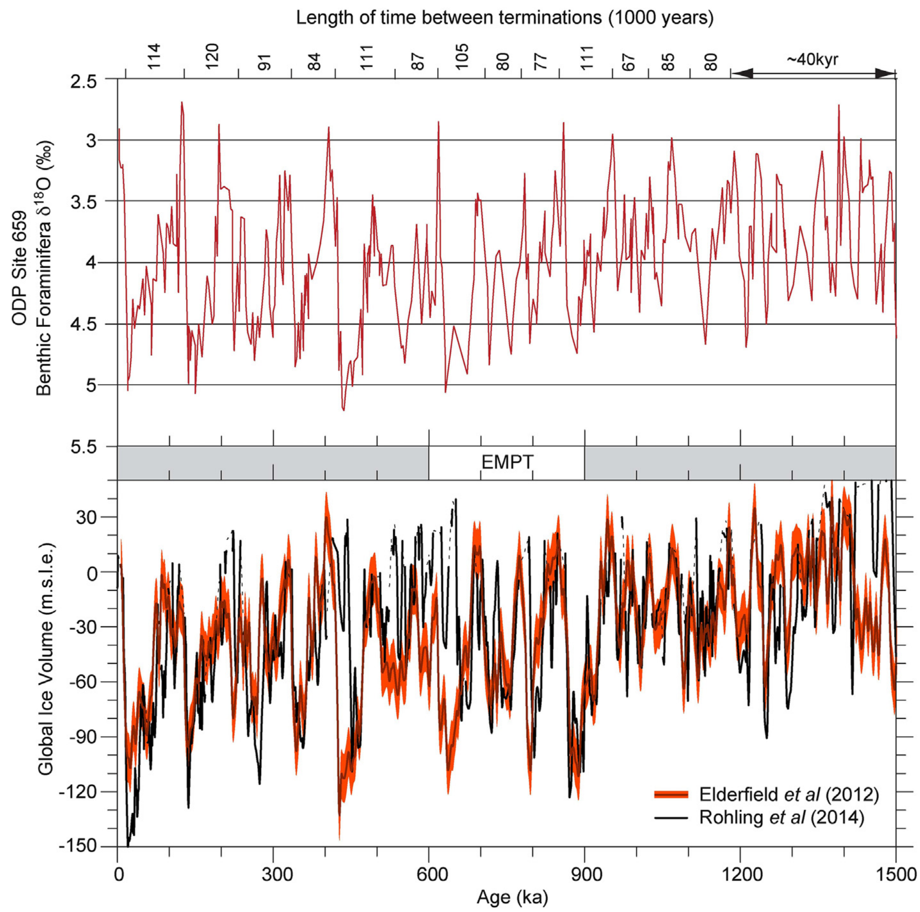

Fig. 1. A) Comparison of the benthic foraminiferal oxygen isotope curve from ODP Site 659 (Tiedemann et al., 1994) with the timing between Terminations (Tiedemann et al., 1994; Raymo, 1997). B) Comparison of two alternate records of global sea level. Black: the marginal sea reconstruction of Rohling et al. (2014) for non-sapropel layers (the dotted portions correspond to the median estimates from linear interpolation and are included for aesthetic reasons). We convert to Global Eustatic sea level, by applying a multiplicative factor of 1.23 chosen for the most recent glacial cycle. Red: Eustatic sea level determined from a deconvolution of combined benthic foraminifera $\delta^{18}O$ and Mg/Ca temperature measurement from the South-west Pacific (Elderfield et al., 2012). We show the mode and 5–95% range from the probabilistic re-interpretaion by Rohling et al. (2014). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

about 900 ka, when there is a significant increase in global ice volume but the 41 ky climate response remains. This situation persists until the second step, about 700 ka, when the climate system finds a three-state solution and strong quasi-100 ky climate cycles begin (Mudelsee and Stattegger, 1997). This is consistent with the more recent evidence from ODP Site 1123 in the Southern Pacific ocean, which shows a step like increase in ice volume during glacial periods starting at MOIS 22 at about 900 ka (Elderfield et al., 2012).

During the EMPT there seems to be a shift from a two stable climate state system to a system with three quasi-stable climate states (Fig. 3). These three states roughly correspond to: 1) full interglacial conditions, 2) moderate glacial conditions such as MOIS 3 that are analogous to the glacial periods prior to the EMPT and 3) maximum glacial conditions for example MOIS 2, the Last Glacial Maximum (LGM). This has also added confused to the definition of the EMPT as many of the intermediate climate periods have been overlooked such as the weak interglacial at ~740 ka, which does not have it own defined MOIS, or the double warm peaks during MOIS 15, 13, and 7.

2. Climate feedback mechanisms

Central to understanding the EMPT is the appreciation that orbital variations do not directly cause global climate changes. Rather they induce small changes in the distribution of insolation across the globe that can in some instances be enhanced by strong positive or negative climate feedbacks and ultimately push the global climate into or out of a glacial period. The initial suggestion by Milankovitch (1949) was that glacial–interglacial cycles were regulated by summer insolation at about 65°N; this was because he reasoned that for an ice sheet to expand additional ice had to survive each successive summer. The focus on the Northern Hemisphere is because the capacity for ice growth is much less in the Southern Hemisphere due to its smaller landmasses combined with the fact that Antarctica is already close to its ice storage limit. The conventional view of glaciation is that low summer insolation in the temperate North Hemisphere allows ice to survive the summer and thus build-up on the northern continents. As snow and ice accumulate the ambient environment is modified. This is primarily by an increase in albedo that reduces the absorption of incident solar radiation, and thus suppresses local temperatures. The cooling promotes the accumulation of more snow and ice and thus a further modification of the ambient environment, causing the so-called 'ice albedo' feedback. Other climate feedbacks such as changes in atmospheric circulation, surface and deep water circulation and the reduction in atmospheric greenhouse gases then play a role in driving the climate into a glacial period (e.g., Berger, 1988; Li et al., 1998; Ruddiman, 2004; Brovkin et al., 2012). These feedbacks then operate in reverse when summer insolation starts to increase (Brovkin et al., 2012; Shakun et al., 2012).