Extended orbitally forced palaeoclimatic records from the equatorial Atlantic Ceara Rise

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Abstract

We extend existing high-resolution Oligocene–Miocene proxy records from Ocean Drilling Program (ODP) Leg 154. The extended record spans the time interval from \(17.86\) to \(26.5\) Ma. The data are age calibrated against a new astronomical solution that affords a re-evaluation of the intricate interaction between orbital ("Milankovitch") forcing of the climate and ocean system, and the fidelity with which this forcing is recorded in oxygen and carbon stable isotope measurements from benthic foraminifera, and associated lithological proxy records of magnetic susceptibility, colour reflectance, and the measured sand fraction. Our records show a very strong continual imprint of the Earth’s obliquity cycle, modulate in amplitude every \(41\) ka, a very strong eccentricity signal in the carbon isotope records, and a strong, but probably local, imprint of climatic precession on the coarse fraction and magnetic susceptibility records. Our data allowed us to evaluate how the interaction of long, multi-million year beats in the Earth’s eccentricity and obliquity are implicated in the waxing and waning of ice-sheets, presumably on Antarctica. Our refined age model confirms the revised age of the Oligocene–Miocene boundary, previously established by analysis of the lithological data, and allows a strong correlation with the geomagnetic time scale by comparison with data from ODP Site 1090, Southern Ocean.

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1. Introduction

Ocean Drilling Program (ODP) Leg 154 data (Curry et al., 1995; Shackleton et al., 1997) so far provide the best developed set of demonstrably astronomically forced lithological and climatic proxy data cycles spanning the Neogene, and extending as far back as the late Eocene (Shackleton et al., 1999). A main strength of palaeo-climate series derived from ODP Leg 154 sediments is that astronomically calibrated age models have been developed based on lithological (Shackleton and Crowhurst, 1997; Weedon et al., 1997; Shackleton et al., 1999) and geochemical (Flower et al., 1997; Pearson et al., 1997) variations, that allow independent geochemical proxies to be placed on a time scale of high accuracy, potentially at the climatic precession period (~21 ka). Multiple studies have exploited the data sets generated from ODP Leg 154 for palaeoceanographic (Shackleton and Crowhurst, 1997; Zachos et al., 1997, 2001a,b; Paul et al., 2000; Flower et al., 2006), biostratigraphic (Rafi, 1999; Shackleton et al., 2000) and cyclostratigraphic (Shackleton et al., 1999; Pälike and Shackleton, 2000; Billups et al., 2004; Pälike et al., 2004; Lourens et al., 2005) purposes, and are now being exploited for the interpretation of data from more recent drilling expeditions (Flowe and Chisholm, 2006). ODP Leg 154 sediments offered the opportunity to study palaeoceanographic processes at a resolution and fidelity previously limited to the Plio- and Pleistocene, over several million years. Some of these studies investigated the close relationship between the cyclostratigraphic record and macroscopic (multi-million year) features of astronomical calculations, made possible by the extended calculations of Laskar et al. (1993). These astronomical calculations have recently been evaluated (Néron de Surgy and Laskar, 1997; Laskar, 1999) and improved (Varadi et al., 2003; Laskar et al., 2004). The age models from ODP Leg 154 were adjusted to these new calculations (Pälike et al., 2004), demonstrating a better fit with these newer solutions compared to the originally used one of Laskar et al.
(1993), and make it opportune to re-visit the impact of this re-tuning on palaeoclimatic proxy series, as well as formally establishing the basis for the re-calibration by publishing the full data series (see supplementary material).

In addition, we extend the original stable isotope record that was established across the Oligocene/Miocene boundary (Paul et al., 2000; Zachos et al., 2001b) both upward and downward, and focus in more depth on the implications of the associated percent sand fraction (>65μm) record, with a series of significant results.

2. Methods

The data sets presented here will be made available in electronic form at a designated data repository. The generation of new benthic stable isotope data (δ18O, δ13C) and percent sand fraction (>65μm) follows the methods described by Paul et al. (2000), and were also measured in Santa Cruz. The initial study by Paul et al. (2000) generated data from two holes: 926B (3′43″N, 42°54′W, 3.6 km water depth) and 929A (5°58″N, 43°44′W, 4.36 km water depth), which were used to generate a spliced record from these two Holes, covering core gaps and identified local slumped horizons. Here, we focused on extending upward and downward the record from ODP Site 926B, the shallower of the two localities, and the one with the higher sedimentation rates. The new record extending downward is based on analyses of a mixture of Cibicidoides species including C. mundulus and C. cresbi, while the new record extending upward is based entirely on analyses of C. mundulus. The complete data we present here were sampled from Core 926B-35X downwards to Core 926B-59X, excluding intervals that were identified as local slumps (Shackleton et al., 1999), and sampled every 10 cm.

The data set also comprised previously generated data from a sub-interval (Paul et al., 2000; Zachos et al., 2001b). Sedimentation rates for this interval were previously estimated by Weedon et al. (1997) and astronomically calibrated by Shackleton et al. (1999): they range between ~25 and 35 m/Ma in the younger time interval, and between ~20 and 30 m/Ma in the older time interval of the record, resulting in a median sample step of ~3.5 ka.

On our new time scale, this entire record now spans the interval ~17.86–26.5 Ma (~17.77–26.24 Ma on the time scales of Shackleton et al., 1999, 2000). The ODP 926B record contains short core breaks, and the data are linked into a composite section by comparison with a spliced combined record of susceptibility and reflectance obtained from several ODP 154 Sites (Shackleton et al., 1999), normalised to zero mean and unit standard deviation as described and used by Pâlîke et al. (2004). The extended data set used in this study is presented in Fig. 1 against Hole 926B core depth in metres composite depth (mcd) (Curry et al., 1995; Shackleton et al., 1999). The stacked susceptibility and colour reflectance record was obtained from multiple Sites, but re-adjusted to the 926B depth scale. It was generated such that positive excursions in this record coincide with positive excursions in the susceptibility record obtained directly from Hole 926B. The coarse fraction data for ODP Hole 926B shown in Fig. 1 supplement those published previously from ODP Site 929 (Zachos et al., 1997; Paul et al., 2000), which had a deeper palaeo-depth. Prior to further analysis, we prepared the percent coarse fraction record by using the transformation arcsin(√%coarse fraction/100). This transformation was applied before further time series analysis in order to achieve more normally distributed values from data that are clipped at zero percent (Weedon, 2003, p. 54).

2.1. Tuning strategies

Significant previous work was undertaken to combine geological data from ODP Leg 154 with calculated variations of the Earth’s orbit in order to achieve an orbitally calibrated age model. This work comprised several important steps. First, sequences from different ODP Leg 154 sites and holes were cross-correlated in order to obtain a continuous record from individual cores, and in order to evaluate the continuity of this “spliced” data set. The correlation was accomplished on the basis of high-resolution lithological data (chiefly, magnetic susceptibility, colour reflectance, down-hole logging data, as well as high-resolution biostratigraphic data). This extensive work, and an evaluation of the stratigraphic continuity of the combined records, was achieved by Weedon et al. (1997) and Shackleton et al. (1999). From these composite spliced records, the next step consisted of analysing the character of any possible astronomical cyclicity on the basis of existing bio-stratigraphic age assignments. Shackleton et al. (1999) used the observation of a dominating obliquity signal and a weaker but discernible climatic precession signal to generate an appropriate astronomical target curve, and used the longer-term amplitude modulation of obliquity (~1.2 million year long amplitude envelope) and climatic precession (~400 ka envelope) to evaluate the fidelity of the generated age model within “hierarchies” of orbital cycle time scales.

The resulting age model, generated on the basis of lithological variations, allowed other proxy data to be placed on this time scale, permitting the calculation of sedimentary fluxes and evaluation of rates of evolutionary change with a resolution and accuracy that is far better than can be achieved by traditional geological means. Shackleton et al. (1999) found that the 1.2 million year long amplitude modulation cycle of obliquity indicated that the age of the Oligocene/Miocene boundary be moved from its previous age of 23.8 Ma (Cande and Kent, 1995) to 22.9 Ma (Shackleton et al., 2000), and that the amplitude variation of climatic precession would place a further constraint on any alternative age model (allowing shifting by ~405 ka multiples). A different view for the correct age of the Oligocene/Miocene boundary was proposed by
Wilson et al. (2002). However, the findings of this study have been questioned (Channell and Martin, 2003), and additional studies confirmed the new correlation of Shackleton et al. (2000) for sediments from other localities (Billups et al., 2004; Pälike et al., 2004), subject to re-adjustment to the new astronomical solution of Laskar et al. (2004).

Importantly, the time scale from Shackleton et al. (1999) was developed using an extension of the calculations that were done by Laskar et al. (1993), which were found to have an inherent limit for the duration over which they can be considered valid into the past (Laskar, 1999). Very recently, an updated and much improved calculation of Earth’s orbital variations was published (Laskar et al., 2004). Pälike et al. (2004) demonstrated using the lithological data from ODP Legs 154 and 199 that this new calculation fits better with the geological data. Here, we report the results of re-adjusting the astronomical age calibration of ODP Leg 154 data to the new astronomical solution, which moves the age of Oligocene/Miocene boundary back in time by ~100 ka, to a new age of 23.03 Ma (Lourens et al., 2005), chiefly controlled by the difference in ~100 ka eccentricity cycle amplitudes.

The re-tuning of magnetic susceptibility data to the new Laskar et al. (2004) solution was done by aligning data and target at the climatic precession frequency, constrained by the ~100 ka amplitude modulation of precession by eccentricity. This assumption of a zero-phase difference at the precession frequency between astronomical solution and geological data results in a phase lag at the obliquity frequency of 6–7 ka during the late Oligocene (Billups et al., 2004). We note here that the new astronomical solution of Laskar et al. (2004) uses current day values for tidal dissipation and dynamical ellipticity (Pälike and Shackleton, 2000), the established phase relationships for obliquity and climatic precession would be altered when different histories for the tidal dissipation and dynamical ellipticity terms were used. A subsequent iteration applied this phase lag at the obliquity frequency for the calculated ETP curve, generating the overall best-fitting target curve.

Fig. 1. ODP Leg 154 data used in this study plotted versus ODP 926B core depth. The ODP 926B record contains short core breaks, and the data are linked into a composite section by comparison with a spliced combined record (a) of susceptibility and reflectance (“SusRef”), normalised to zero mean and unit standard deviation as described and used by (Pälike et al., 2004). Also plotted are (b) the colour reflectance parameter L* (lightness) and (c) magnetic susceptibility (“SUS”), calculated by gaussian interpolation at stable isotope sample points. Previously published benthic foraminiferal data from ODP Site 926B (Paul et al., 2000; Zachos et al., 2001b) were extended and are plotted with (d) the >63 µm coarse fraction data. Benthic δ13C (e) and δ18O (f) are plotted referenced against VPDB. Mi–1 denotes the position of a prominent glacial event previously discussed (Miller et al., 1991; Zachos et al., 2001b).
The zero-phase assumption at precession stems from the evidence that the strong precession signal arises from local climatic processes on the adjacent continent which modulate the terrigenous input, and possibly local productivity and the preserved sediment coarse fraction. The lag at the obliquity frequency presumably arises from the slow response of the varying Antarctic continental ice sheet, and for δ13C on the long residence time of carbon and nutrients in the ocean. The phase relationship used follows that of Shackleton and Crowhurst (1997). They interpreted magnetic susceptibility values as a proxy for the terrigenous supply and correlated maxima in susceptibility with maxima in Southern Hemisphere summer insolation (minima in Northern Hemisphere insolation), based on the observations that otherwise positive δ18O values would correspond to Northern Hemisphere insolation minima, and the obliquity response would be out of phase. In this scenario, increased terrigenous input through the Amazon system would be driven by maxima in Southern Hemisphere insolation. They also argued, though, that in the Oligocene the uncertainties in the orbital calculations are still at an order (Pa¨like and Shackleton, 2000; Lourens et al., 2000) where age errors that arise from such phase assumptions are smaller than the error that arises from our incomplete knowledge of the dissipative effects (tidal dissipation) in the orbital solutions. The data sets are plotted against our new time scale in Fig. 2, together with the astronomical templates from Laskar et al. (2004), as well as a comparison of the previous time scale of Shackleton et al. (1999) that was generated with the astronomical solution by Laskar et al. (1993) and applied by Zachos et al. (2001b) to a sub-set of the stable isotope data. “ETP” refers to an artificial mix of eccentricity, obliquity (tilt) and climatic precession, that are combined by normalising these three components (subtracting their mean and dividing by their standard deviation), and then adding these with a set of weightings determined by the geological data. In this case the eccentricity component was accentuated as the stable isotope data have a large eccentricity imprint. We note that ETP resembles Northern

![Fig. 2. ODP Leg 154 data plotted versus revised astronomical age (Laskar et al., 2004). Shown are (a) the previously published spliced δ18O dataset from Sites 926 and 929, age calibrated using a previous astronomical solution (Zachos et al., 2001b; Laskar et al., 1993), against new extended δ18O (b) and δ13C (c) data, using a revised astronomical solution and time scale (Laskar et al., 2004). (d) Shows an arcsin(√%) transformation of the age-calibrated coarse fraction percentage data. This transformation was applied before further time series analysis in order to achieve more normally distributed values from data that are clipped at zero percent (Weedon, 2003, p. 54). (e) Shows age-calibrated combined susceptibility and lightness data. (f) Shows Earth’s obliquity and ~1.2 Ma obliquity amplitude modulation from Laskar et al. (2004). (g) Calculated mix of eccentricity, obliquity (tilt), and climatic precession (“ETP”), using Laskar et al. (2004), with relative amplitudes reflecting those of the data, and ~6.4 ka lagged obliquity component (Billups et al., 2004). (h) 400 ka bandpass filters (±0.5 Ma−1 gaussian bandwidth) applied on ETP (black), −δ18O (blue) and −δ13C (green). Red tick marks and codes mark the position, count and identification of the absolute number of ~400 ka cycles as first proposed by Wade and Pa¨like (2004). Thin cross-lines mark the position of obliquity amplitude minima, thick green cross-lines mark the position of long-term eccentricity amplitude minima.](image-url)
Hemisphere summer insolation maxima at climatic precession minima, depending on the definition of the climatic precession phase angle (Berger et al., 1993). In Fig. 2(h) we have also marked the position of absolute ~405 ka cycle codes, as first described by Wade and Pälike (2004), supplementing the code with a letter abbreviation for the geological epoch (Mi for Miocene, Ol for Oligocene) and an indication within which magnetic reversal interval this cycle falls for existing astronomically calibrated sections (Lourens et al., 2005), following the convention of Cande and Kent (1995).

2.2. Spectral analysis

We conducted different spectral analysis methods on the astronomical curves, the benthic stable isotope data, the coarse fraction record, and the spliced and stacked susceptibility and reflectance ("SusRef") record. The interval analysed extends from 17.86 to 26.5 Ma. Prior to analysis, the data were re-sampled at the median sample resolution every 4 ka. Data sets were linearly detrended, and long-term trends (periods longer than ~1 Ma) were removed with a gaussian notch filter, using version 2 of the software package AnalySeries (Paillard et al., 1996). Prior to analysis, we multiplied the two stable isotope series with temperature minima (or ice-volume maxima). We used three main tools for spectral analysis. Power spectra were calculated using the Multitaper method with a timestep bandwidth product of 4 and 6 tapers, using the SSA-MTM Toolkit v4.2 (Ghil et al., 2002). This software was also used to fit red-noise significance limits at the 90% and 99% levels. We used the Blackman–Tukey method in AnalySeries to calculate cross-spectral coherence and phase relationships between an artificial mix of eccentricity, obliquity (tilt), and climatic precession (ETP). For the coherency, we determined the 6 tapers, using the SSA-MTM Toolkit v4.2 (Ghil et al., 2002). This software was also used to fit red-noise significance limits at the 90% and 99% levels. We used the Blackman–Tukey method in AnalySeries to calculate cross-spectral coherence and phase relationships between an artificial mix of eccentricity, obliquity (tilt), and climatic precession (ETP). For the coherency, we determined the 95% confidence interval testing the non-zero hypothesis. We also computed evolutive Multitaper-method spectra for the data, using a sliding 1.6 Ma window stepped at 50 ka intervals. For the cross-spectral analysis the resulting frequency bandwidth is 0.86 Ma⁻¹.

3. Results

3.1. ODP Leg 154 data re-calibrated to new astronomical calculation

Our first result concerns the re-adjustment of the ODP Leg 154 data set to the refined astronomical solution of Laskar et al. (2004). Fig. 2 (a) and (b) shows a comparison of the benthic oxygen isotope data as used by Zachos et al. (2001b), using the time scale of Shackleton et al. (1999), as well as the re-adjusted data on the time scale of our study. In terms of the Shackleton et al. (1999) time scale, our new ages are approximately 20 ka older in the interval prior to ~19.71 Ma (using Shackleton et al., 1999 ages). This age shift towards slightly older ages then gradually increases between ~19.71 and 20.15 Ma and reaches an average offset of approximately 90 ka for the remainder of the data, except one interval between ~23.35 and 24.21 Ma, where this reaches ~160 ka. This difference is chiefly controlled by the different nature of short eccentricity cycles in the two different target curves used (for the offset of ~90 ka or one eccentricity cycle), and some additional small refinements in the original splice that were done after the publication of Shackleton et al. (1999) (N. J. Shackleton, pers. comm. 2002). We note that on our new re-adjusted time scale, evaluated in detail by Pälike et al. (2004), the position of the oxygen and carbon isotope excursions previously designated as “Mi–1” again fall into a “node” of minimum obliquity amplitude variation, and coincide approximately with the position of the Oligocene/Miocene boundary (Shackleton et al., 2000). The revised ages presented in our study were the basis for the age given for the O/M boundary in Lourens et al. (2005). Some of the additional magnetic reversal ages given in Lourens et al. (2005) were derived from Australia–Antarctic spreading distances, which are said to improve the original approach of fitting a smooth spline curve to a smaller number of tiepoints as was done by Cande and Kent (1995). The fact that Mi–1, here designated as long eccentricity cycle 58, falls into an obliquity amplitude node, is consistent with the original placement in Zachos et al. (2001b). However, the overall influence of the orbital geometry on glaciations is more complex than previously argued, and will be discussed below.

3.2. Time evolution of spectral characteristics

The combination of a refined astronomical age scale together with an extension of the time coverage of the data allows a re-evaluation of the spectral characteristics contained in the data sets, and an enhanced analysis of the relationship between long-term astronomical forcing and the encoding of this forcing in the different data sets. The spectral characteristics of the data and the astronomical solution of Laskar et al. (2004) are summarised in Fig. 3. For all data sets, the ~40 ka obliquity cycle is the one that most significantly stands out above the background spectrum. This is the reason why the age model of Shackleton et al. (1999) used this cycle as the pre-dominant tuning target, in combination with the climatic precession imprint in the SusRef data set. Both stable isotope time series contain significant short and long eccentricity cycles (~110 and ~405 ka). Interestingly, of the palaeoceanographic proxy time series, the coarse fraction record contains significant spectral peaks at the three climatic precession frequencies, which were also established in the spliced record of Zachos et al. (2001b). We can now re-evaluate the fidelity of the ~1.2 Ma long obliquity amplitude modulation cycle that was already established for the SusRef record (Shackleton et al., 1999; Pälike et al., 2004). Throughout the record, the oxygen isotope data
display obliquity cycle amplitude variations co-varying with the astronomical target curve. Minima in the amplitude of the ~41 ka cycle of the astronomical data are indicated by dark shaded lines in Fig. 3. This pattern also fits for the benthic carbon isotope record. The evolutive spectrum of the coarse fraction record also displays the ~2.4 Ma long amplitude modulation of eccentricity at the climatic precession frequency bands. This feature is also apparent in the amplitude variation of the ~405 ka long eccentricity cycle for the benthic carbon isotope data. Phase relationships between astronomical and geological data are similar to previous results (Zachos et al., 2001b), as shown in Fig. 3(c) and (f). The climatic precession component of the tuning data set SusRef is in
phase with the astronomical target. The obliquity is in phase between the lithological data set and the coarse fraction record with respect to lagged obliquity. The benthic oxygen isotope data show a further lag of about 3–4 ka at the obliquity frequency, while carbon isotope data show a further lag of about 180° out of phase with the oxygen isotope data. This is in contrast to the eccentricity periods, where oxygen and carbon isotopes co-vary, and also to observations from other localities spanning the same time interval (Billups et al., 2004). In terms of absolute phase lags (in ka), it is interesting to note that the lag of the carbon isotopes in particular increases with increasing cycle periods, an observation that is compatible with the predictions of box models of the carbon cycle if forced by astronomy, specifically with respect to the size of the carbon reservoir and its preferential filtering of longer periods due to the residence time of carbon in the oceans (Pälike, unpublished results, using a modified version of the Walker and Kasting, 1992 model).

3.3. Comparison with Southern Ocean record and global compilation

We can now compare our extended tuned records with one that was recently developed for a benthic stable isotope data set from ODP Leg 177, Site 1090, Southern Ocean, Atlantic sector (Billups et al., 2002, 2004), which also used the new astronomical solution of Laskar et al. (2004) as a tuning target. The comparison of our data set with that from the Southern Ocean is shown in Fig. 4. The record from ODP Leg 177 contained a well-defined data set of magnetic polarity measurements (Channell et al., 2003), and the ages obtained for ODP Leg 154 can be used to refine the geomagnetic polarity time scale (Cande and Kent, 1995). The age models for our present study and that of Billups et al. (2004) were obtained independently, yet show a close match, particularly of the carbon isotope record (Fig. 4). As discussed by Billups et al. (2004), across the Oligocene/Miocene interval the basin-to-basin offset of carbon isotope values is virtually absent, even 5 million years into the Miocene. The Southern Ocean oxygen isotope values show a consistent heavy offset of ~0.5‰ with respect to the equatorial Atlantic values. Fig. 4 also shows a comparison of our extended data set with the global multi-site compilation of Zachos et al. (2001a). The original data set from Zachos et al. (2001b) made up a large portion of the global compilation curve back to ~25 Ma on our new time scale. As shown by Lear et al. (2004) using a new data set from the equatorial Pacific, the apparent “late Oligocene warming” that is suggested by the global compilation is largely artefact of using ODP Leg 154 data for ages younger than ~25 Ma, and sub-Antarctic data for the older part of the Oligocene. The stable isotope data presented here show a much more gradual and smaller decrease towards lighter values over more than 4 million years before the onset of Mi–1. We provide a close-up of the isotope data sets in Fig. 5, which compares the independently age-calibrated data sets from ODP Site 1090 (Billups et al., 2004), ODP Hole 929A (Zachos et al., 2001b) and our study that uses data from ODP Hole 926B. In most cases the oxygen and carbon isotope data show that the three records agree down to the obliquity scale level of variations. The new astronomical solution La2004 (Laskar et al., 2004) does still exhibit a node in the obliquity amplitude variation, but this is not as pronounced during the Mi–1 event as previous astronomical solutions. Thus, in order to explain the magnitude of Mi–1 compared to previous and following obliquity nodes a further pre-conditioning, for example through a decline of atmospheric pCO2 concentrations at this time is required (Deconto and Pollard, 2003).
3.4. Evaluation of amplitude modulations

Our new age model suggests a re-evaluation of the astronomical signature in the geological data as encoded by the amplitude modulation terms of obliquity and eccentricity (Laskar, 1999; Shackleton et al., 1999; Pâlike et al., 2004). The results of complex demodulation at the obliquity frequency are presented in Fig. 6(a). This analysis shows that the \( C24^1:2 \) Ma amplitude modulation of the \( C24^4 \) ka obliquity signal is present, and approximately in phase with the astronomical solution, for all four time series analysed. The results here also confirm the analysis by Pâlike et al. (2004). We also computed the amplitude variation at the climatic precession frequency, which is modulated by the short and long eccentricity, as well as a longer \( C24^2 \) Ma long term (Laskar, 1999). Here, we extracted the \( C24^405^1:2 \) ka amplitude modulation term, as presented in Fig. 6(b). Apart from an interval around 20 Ma the data modulation is again in phase with the astronomical modulation, and the data sets also suggest the presence of the longer multi-million year modulation between \( C24^21 \) and \( C24^23.8 \) Ma. We did not analyse the climatic precession frequency for the stable carbon isotope data as this signal is not significantly above the noise background (Fig. 3(a)), presumably due to the low-pass filtering nature of the oceanic carbon reservoir.

3.5. Enhanced variability of carbon isotope data prior to ca. 25.75 Ma

The carbon isotope data show enhanced variability before around 25.75 Ma (Figs. 2 and 4). At first sight, these data look like a larger noise contribution, and their local nature is suggested by the absence of this larger variability in records from the Pacific (Pâlike et al., unpublished data). However, on closer inspection the variability cannot simply be explained by individual data point outliers. Instead, the enhanced variability appears to closely follow an astronomical forcing (Fig. 7). The analysis shows a strong obliquity imprint in the >1‰ \( \delta^{13}C \) variations. In contrast to the remainder of the data set, here the phase relationship between carbon isotopes and astronomical obliquity is reversed, such that light carbon isotope values correspond to low obliquity and insolation. There are two possible
explanations for the low values: (1) a “Mackensen” (Mackensen et al., 1993) effect possibly associated with more frequent algal blooms, or (2) systematic shifts in Cibicidoides species abundances towards those with more depleted values. In this interval, because of the lack of C. mundulus, species were mixed to generate sufficient carbonate for isotope analysis. Either of these artefacts could be related to local responses to changes in astronomical forcing (i.e., the strength of the trade winds). In addition, δ\textsuperscript{13}C minima coincide with coarse fraction minima over this interval, indicating the possibility of an enhanced influx of corrosive water.

4. Discussion

4.1. Glaciation events and orbital confluences

We previously suggested that the Mi–1 glaciation event at 23 Ma was related to a rare confluence of reduced obliquity amplitude variations (a obliquity “node”) in combination with low eccentricity (Zachos et al., 2001b). The new data presented here as well as the new astronomical solution used allow us to refine our previous findings. First, in the new astronomical solution, there is still an obliquity node at 23 Ma, however, it is not as pronounced as it was in the previous astronomical calculation of Laskar et al. (1993). In the new solution of Laskar et al. (2004), the most extreme obliquity nodes occur at around 19.7, 20.7, 24.4 and 25.4 Ma (Fig. 2). Our age model is sufficiently robust to exclude the possibility that the age for Mi–1 has to be re-assigned to either 24.4 or 20.7 Ma. At most obliquity nodes (but not all) we do indeed observe isotope excursions toward more heavier values in δ\textsuperscript{13}C and δ\textsuperscript{18}O, however, our data suggest that the

Fig. 6. Comparison of amplitude modulation between astronomical solution and ODP 154 data: (a) obliquity amplitude complex demodulation (Bloomfield, 1976) implemented using the software “TRIP” (Gastineau and Laskar, 2005). The data and astronomical solution were demodulated at a period of ~40.17 ka, and from the demodulation the ~1.2 Ma amplitude modulation was extracted, and normalised to standard deviation and zero mean. (b) Climatic precession/eccentricity amplitude complex demodulation. The data and astronomical solution were demodulated at a period of ~18.8 ka, and from the demodulation the ~400 ka amplitude modulation was extracted, and normalised to standard deviation and zero mean.

Fig. 7. Close-up view of benthic δ\textsuperscript{13}C between 25.5 and 26.5 Ma. Top panel shows carbon isotope data (dashed green) and astronomical “ETP” curve for the interval that shows unusually strong carbon isotope fluctuations (see Figs. 1 and 2). Middle panel shows these two time series after applying a gaussian bandpass filter centred around the climatic precession frequencies (47 ± 9 Ma\textsuperscript{−1}). Bottom panel shows these two time series after applying a gaussian bandpass filter centred on the main obliquity frequency (25 ± 5 Ma\textsuperscript{−1}). Results suggest that strong local δ\textsuperscript{13}C fluctuations are at least partly driven by a local process.
link between astronomy and climatic excursions is more intricate than previously suggested, and agrees well with the finding by Wade and Pälike (2004) from an earlier part of the Oligocene that glacial conditions are most readily observed when the climate system is pre-conditioned by low obliquity amplitude variations, and then triggered by individual ~110 and 405 ka eccentricity extremes. This relationship is now also found for the main mid-Miocene Mi–3b cooling event (Abels et al., 2005; Holbourn et al., 2005). Existing and new data from the Pacific suggest that the entire Oligocene was globally dominated by a strong ~405 ka cycle, particularly in the benthic isotope record, and that glaciations are typically enhanced during most obliquity nodes. For example, the extreme glaciation event Oi–1, during long 405 ka eccentricity cycle 84 OlC–13n, was also pre-conditioned by an obliquity node (Coxall et al., 2005). Further modeling studies are in progress that will elucidate the link between astronomical forcing of the oceanic carbon cycle and the carbon and oxygen benthic isotope records. Support for the view that the strength of coarse fraction variations co-evolves with the obliquity amplitude (Paul et al., 2000, our Fig. 6(a)).

4.2. Evolution of coarse fraction record and astronomical forcing

An interesting observation is that our coarse fraction record from Hole 926B shows a much stronger response at the climatic precession frequencies than any of the other data sets (Fig. 3(d)), although the signal at eccentricity frequencies, which should modulate the strength of the precession cycle, is much weaker. We interpret the strength of the climatic precession cycle in the coarse fraction record as a local rather than global feature, as judged by the much weaker recording of this signal in the stable isotope records. As argued by Zachos et al. (1997), the coarse fraction signal can be influenced by three factors: (1) changes in carbonate production, foraminiferal size variations and or flux rates of coccolithophores related to the intensity of the trade winds; (2) changes in dilution of coarse fraction by varying influxes of clays from the near Amazon (Shackleton and Crowhurst, 1997) and (3) varying amounts of dissolution. Zachos et al. (1997) argued for (3), as they observed the general association of a small coarse fraction with increased evidence of fragmentation, and lower carbonate content as indicated by higher susceptibility measurements (see Fig. 3(f)). While the strong additional signal of obliquity in the coarse fraction record could suggest that at least part of the signal is contributed by (globally) increased circulation of corrosive bottom waters, the phase difference between δ18O and coarse fraction is significant enough to suggest that this was not the only contributing factor. Further studies will be needed to de-convolve the relative contributions of the three possibilities above.

4.3. Inter-site comparison of ODP 926 and 929 coarse fraction data

A previous study (Paul et al., 2000) published a coarse fraction record (> 65μm) for Site 929, which can be compared to the record from the shallower Site 926 (Fig. 8). The deeper Site 929 shows a reduced coarse fraction content, compatible with increased effects of dissolution and fragmentation of foraminifera with depth. For Site 929 in particular, there is a marked decrease in coarse fraction for sample ages younger than ~22 Ma. We observe that individual coarse fraction maxima are

Fig. 8. Comparison of previously published coarse fraction record from Site 929 (Paul et al., 2000) and the record from the shallower Site 926. Also shown are Earth’s obliquity and the difference in > 65μm % coarse fraction between the two sites. The record from the deeper Site 929 shows a smaller coarse fraction, as expected if this process was dissolution controlled. Both coarse fraction records are in phase at obliquity periods. Larger differences between the two records are also in phase with high obliquity values. In addition, there appears a discernible trend of enhanced coarse fraction amplitude at ~1.2 Ma obliquity amplitude maxima, as well as an increase of coarse fraction site-to-site differences at ~1.2 Ma obliquity amplitude minima. Vertical dashed lines are inserted to aid comparison of the phasing of individual peaks. All plotted data are interpolated at 5 ka time steps.
approximately in phase with obliquity maxima, consistent with the observation that glacial maxima are correlated with obliquity maxima, and with a higher carbonate ion concentration and deeper lysolcine. We also calculated the difference between the age interpolated records (Site 926 minus Site 929 coarse fraction). As a general trend, we observe that this indicator of depth-dependent dissolution (i.e., a deeper lysolcine) decreases during maxima of the ~1.2 Ma long obliquity amplitude modulation. In the upper half of the record, we also observe that the obliquity cycles manifested in the coarse fraction records from the two sites have not cancelled out after taking the difference, partly due to coarse fraction that is close to zero for Site 929.

5. Summary

The extension of the original data set from Zachos et al. (1997) presented here, in combination with a refined astronomically calibrated age model, provides a standard reference section for a time during Earth’s history where periodic glacial advances and retreats on Antarctica were extremely sensitive to variations in the boundary conditions, such as astronomical insolation forcing. The records presented here provide a degree of chronostatigraphic control around the Oligocene/Miocene transition that was previously exclusive to the Quaternary. Pioneering the combination of astronomical age models by N. J. Shackleton with very high-resolution stable isotope measurements is one of the strongest contributions and turning points in the Earth System Science that only now are we beginning to exploit to its full extent.

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