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Estimate of climate sensitivity from carbonate microfossils dated near the Eocene-Oligocene global cooling

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Received: 9 September 2012 – Accepted: 17 September 2012 – Published: 5 October 2012

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Published by Copernicus Publications on behalf of the European Geosciences Union.

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Abstract

Climate sensitivity is a crucial parameter in global temperature modelling. An estimate is made at the time 33.4 Ma using published high-resolution deep-sea temperature proxy obtained from foraminiferal $\delta^{18}\text{O}$ records from DSDP site 744, combined with published data for atmospheric partial pressure of CO_2 ($p\text{CO}_2$) from carbonate microfossils, where $\delta^{11}\text{B}$ provides a proxy for $p\text{CO}_2$. The $p\text{CO}_2$ data shows a $p\text{CO}_2$ decrease accompanying the major cooling event of about 4°C from greenhouse conditions to ice-cap conditions following the Eocene-Oligocene boundary (33.7 My). During the cooling $p\text{CO}_2$ fell from 1150 to 770 ppmv. The cooling event was followed by a rapid and huge increase in $p\text{CO}_2$ back to 1130 ppmv in the space of 50 000 yr. The large $p\text{CO}_2$ increase was accompanied by a small deep-ocean temperature increase estimated as $0.59 \pm 0.063^\circ\text{C}$. Climate sensitivity estimated from the latter is $1.1 \pm 0.4^\circ\text{C}$ (66% confidence) compared with the IPCC central value of 3°C . The post Eocene-Oligocene transition (33.4 Ma) value of 1.1°C obtained here is lower than those published from Holocene and Pleistocene glaciation-related temperature data (800 Kya to present) but is of similar order to sensitivity estimates published from satellite observations of tropospheric and sea-surface temperature variations. The value of 1.1°C is grossly different from estimates up to 9°C published from paleo-temperature studies of Pliocene (3 to 4 Mya) age sediments. The range of apparent climate sensitivity values available from paleo-temperature data suggests that either feedback mechanisms vary widely for the different measurement conditions, or additional factors beyond currently used feedbacks are affecting global temperature- CO_2 relationships.

1 Introduction

Estimation of climate sensitivity to greenhouse gas concentration from the geological past is a useful tool which assists in the assessment of the risks of future climate change associated with current anthropogenic emissions of greenhouse gases. For

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geological records the available data is largely restricted to CO₂ since other greenhouse gases are not embedded in the geological record with sufficient clarity. The task of separating influences of tectonic change (affecting mountain building and oceanic pathways) is increasingly difficult as one searches for suitable data sets further back in time. That separation is easier when a change in atmospheric CO₂ concentration can be shown to occur in a (geologically) short period of time since this gives opportunity to observe the impulse response of temperature change associated with a rapid atmospheric CO₂ change. A coincidence of paleo-temperature and CO₂ variations in the geological record does not necessarily prove cause and effect, and unquestionably the relationship will be multifactorial, however any observed relationship does provide some bounds on possible models for the relationship.

In this paper we consider an event in the early Oligocene (from 33.2 to 33.5 Ma) following the Eocene-Oligocene transition (EOT) where a high resolution deep ocean temperature record is combined with estimates of atmospheric CO₂ to give such an impulse response.

2 An estimate of short-term temperature shift after the EOT

We use $\delta^{18}\text{O}$ records from benthic foraminiferas acquired in Deep Sea Drilling project (DSDP) holes 744 and 522 as published by Zachos et al. (1996, hereafter referred to as ZQS). Hole 744 is at latitude 60° S on the Kerguelen Plateau off the coast of Antarctica, at an ocean depth of 2307 m. Hole 522 is at latitude 25° S in the Cape Basin off the coast of Namibia, at an ocean depth of 4444 m. These holes provide $\delta^{18}\text{O}$ records with a resolution of order 10 000 yr across the Eocene-Oligocene boundary thus providing an excellent proxy for deep-ocean temperature.

Figure 1 shows the raw data for hole 744, together with a smoothed plot (used for visual inspection only). The data shows the increase in $\delta^{18}\text{O}$ associated with major cooling from the late Eocene to early Oligocene. ZQS describes this cooling as being a change from terrestrial greenhouse conditions with no ice-caps, to glacial conditions

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probably with a single (south) polar icecap having dimensions about 40 % of the current Antarctica ice-cap. Calculation of the global temperature change from the $\delta^{18}\text{O}$ data is a function of both ocean temperature and ice volume and is estimated by ZQS to be in the range 3 to 4 °C. Following the temperature drop to a minimum at 33.4 Ma a step change of duration about 150 000 yr is evident on the smoothed data. The raw data shows spiky variations which Zachos et al. show from time-series analysis to have a dominant period of 41 000 yr corresponding to glacial cycles at the Milankovich tilt variation frequency.

The step change in $\delta^{18}\text{O}$ is quantified by making some subjective choice on the position of time segments labelled a, b, c on Fig. 1, selected as representative samples before, during and after the temperature change associated with the CO_2 pulse shown in Fig. 4. Having selected the time intervals, average values and uncertainty for $\delta^{18}\text{O}$ can be calculated as shown in Table 1. Figures 2 and 3 show corresponding data for hole 522, where records for two different foram species are plotted separately as they show different baselines in the $\delta^{18}\text{O}$ values. Data for hole 522 is subject to much higher uncertainty, partly a result of missing data within the time segments a and c. Averages shown in Figs. 2 and 3 are not regarded as meaningful for the calculation of a climate sensitivity value in this paper, but are included in Table 1 for completeness.

From Table 1 using data from hole 744 we have

$$\Delta\delta^{18}\text{O} = 0.16 \pm 0.018\text{‰} \quad (1)$$

Using the relationship between isotopic partition between carbonate and water, as a function of temperature T_{water} , as established by Shackleton (1974) and approximated by Duplessey (2002),

$$(\delta^{18}\text{O}_{\text{CO}_3} - \delta^{18}\text{O}_{\text{water}}) = 0.28 T_{\text{water}}, \text{ for } T \sim 0^\circ\text{C} \quad (2)$$

which gives for the deep ocean at hole 744

$$\Delta T_{\text{water}} = 0.59 \pm 0.063^\circ\text{C} \quad (3)$$

assuming constant polar ice volume and consequent constant $\delta^{18}\text{O}_{\text{water}}$. Variations in $\delta^{18}\text{O}_{\text{water}}$ are known to account for in excess of 50 % of $\delta^{18}\text{O}_{\text{CO}_3}$ in variation in deep Pleistocene glaciations (de Boer, 2010) and are estimated to account for 30 % of the $\delta^{18}\text{O}_{\text{CO}_3}$ change across the EOT (ZQS). For the small variations in temperature and ice volume associated with the $p\text{CO}_2$ pulse post EOT, the 30 % figure can be considered an upper limit. If the 30 % figure were to be applied, then the ΔT_{water} estimate in Eq. (2) reduces by a factor 0.7.

The relationship between deep-ocean temperature and mean global temperature is not documented in Oligocene geological records but guidance from sediment and ice-core records from Pleistocene glaciation gives a guide. While tectonic changes and associated gross ocean current changes are expected to influence such relationships over geologic time, the fact that the separation of the South America and Antarctica land masses had occurred prior to the post-EOT time under discussion is important. The opening up of the ocean permitting a circum-polar ocean current, and the geologically very short duration of the time under discussion, mean that the likelihood of perturbations of temperature relations post EOT being very different from Pleistocene values is much reduced.

Kohler et al. (2010, their Fig. 8) show Pleistocene variations in North Atlantic deep ocean temperatures closely tracking estimates of variations in global temperatures, for the warmer halves of the past eight Pleistocene glacial cycles (the variations do not correlate for the colder halves of the cycles since ocean water temperatures cannot reduce below about -1°C). Since global temperatures in the post EOT time under discussion are approximately equivalent to, or may be a degree or so warmer than, peak interglacial temperatures (with a unipolar ice-cap as inferred by ZQS), a linear relation between deep ocean and global temperatures is a reasonable assumption for the post EOT. An alternative relationship based on a similar study of Pleistocene glaciations by Hansen and Sato (2012) suggests global average temperature is a factor 1.5 greater than deep ocean temperature. We thus use a value for global average temperature deduced from Eq. (3) to be

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$$\Delta T_{\text{global}} = S_1 S_2 (0.59 \pm 0.063) ^\circ\text{C} \quad (4)$$

where S_1 is in the range 1.0 to 0.7 (ice volume correction), and S_2 is in the range 1.0 to 1.5 (ratio of deep ocean to global temperature variation for inter-glacial times).

3 An estimate of a pulse of atmospheric $p\text{CO}_2$

5 We use estimates of atmospheric CO_2 concentration ($p\text{CO}_2$) from Pearson et al. (2009, referred to hereafter as PFW) obtained by study of $\delta^{11}\text{B}$ isotopes in upper-ocean planktonic foraminifera preserved in the Kilwa formation of coastal Tanzania. Figure 4 and Table 1 show $p\text{CO}_2$ values computed as an average of two low and two high values of $p\text{CO}_2$ in the PFW data set. Uncertainty of the average values and the difference are
10 computed using 1-sigma uncertainties (being half the 2-sigma uncertainties provided by PFW).

Thus we have

$$\Delta p\text{CO}_2 = 358 \pm 101 \text{ ppmv} \quad (5)$$

from a baseline of 774 ppmv.

15 It is obvious that there is some discrepancy in timing between the time segments for the $\delta^{18}\text{O}$ and the $p\text{CO}_2$ data. The relationship between temperature and $p\text{CO}_2$ is in this example is discussed quantitatively by PFW who used results of Merico et al. (2008) to conclude that the $p\text{CO}_2$ increase and decrease following the EOT are consistent with carbon-cycle models, apart from a discrepancy that the models predict the $p\text{CO}_2$
20 rebound to occur over 500 000 yr whereas the observations indicate the rebound occurred over 50 000 yr.

For the purposes of this study we take the empirical view that the $p\text{CO}_2$ pulse occurred, and associated with the pulse there was a measureable deep-ocean temperature shift. We therefore use the two parameters to place some limits on climate

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sensitivity (CS) at the time. The imperfect synchronization of the start of the temperature increase and the start of the $p\text{CO}_2$ rebound adds to uncertainties, but a value for CS near the EOT is a useful addition to paleo-climate records.

4 An estimate of climate sensitivity

5 Computing climate sensitivity for a doubling of $p\text{CO}_2$ gives from Eqs. (4) and (5),

$$\begin{aligned} \text{CS}_{2\times\text{CO}_2} &= \Delta T_{\text{global}} / \log_2(1 + \Delta p\text{CO}_2 / \text{Base}_p\text{CO}_2) \\ &= 1.1^\circ\text{C}, \text{ range } 0.7 \text{ to } 1.5; 66\% \text{ confidence.} \end{aligned} \quad (6)$$

This estimate hereafter called CS_{EOT} is also subject to systematic errors associated with the uncertainty of scale-factors S_1 and S_2 noted at Eq. (4).

10 5 Significance relative to other recent estimates of climate sensitivity

It is reasonable to classify values of CS obtained from paleo-temperature data as being for equilibrium climate sensitivity, incorporating both fast and slow feedbacks, whereas values obtained from satellite or historical meteorological data represent transient climate sensitivity, incorporating fast feedbacks only. Lunt et al. (2010), Rohling et al. (2012) and Hansen and Sato (2012) provide discussions on the differences.

15 CS for a “no-feedback model” can be computed for an earth with atmosphere exhibiting infra-red absorption by CO_2 but no additional feedbacks associated with water-vapor, clouds and aerosols. The value is given as about 1°C (e.g. Lindzen and Choi, 2011) or 1.2°C (Hansen and Sato, 2012).

20 The value for CS_{EOT} obtained here is compared in Table 2 with values published from other methodologies. The value obtained is similar to values for a “no-feedback model”, and is comparable with CS obtained from satellite data (Douglas and Christy, 2009; Lindzen and Choi, 2011) and is placed at or below the low end of ranges of CS

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obtained from meteorological data by Annan and Hargreaves (2011), Lewis (2012), and Gillett et al. (2012). It is significantly below the ranges given by the IPCC (Solomon et al., 2007) and Forest et al. (2006).

Among values obtained for CS from Pleistocene and Holocene paleo-temperature data, CS_{EOT} is at or below the low end obtained by Chylek and Lohmann (2008), Kohler et al. (2010) and Schmittner et al. (2011). It is significantly below values presented by Lea (2004), Rohling et al. (2012) and Hansen and Sato (2012).

Values of CS obtained from Pliocene paleo-temperature data (Pagani et al., 2010; Dowsett et al., 2012), shown in Table 2, are much higher than CS_{EOT} by a factor 6 to 8. There is a quite fundamental difference in the nature of the experimental system studied for the Pliocene examples; the atmospheric CO_2 is in a steady or slowly-varying state over time periods of millions of years, rather than being an impulsive increase as with the post EOT example.

An obvious comparison of the post EOT CO_2 pulse is with the pulse at the Paleocene-Eocene Thermal Maximum (PETM) (55 Ma). Zeebe et al. (2009) describe the event as having a temperature increase of 5 to 9 °C over a few thousand years, accompanied by the release of about 3000 Pg of carbon into the atmosphere, producing a pCO_2 increase from about 1000 ppmv to 1700 ppmv. The large temperature change is a factor 2 to 5 times that which might be predicted using the IPCC CS, and Zeebe et al. (2009) conclude that feedbacks and/or forcings other than atmospheric CO_2 caused a major portion of the PETM warming. The post EOT CO_2 pulse discussed in this paper is likewise very large (about 50 % of the magnitude of the PETM pulse), but is accompanied by a warming only about a factor 0.3 of that expected from the IPCC central value of CS. This suggests the feedbacks and forcings for the two events are very different.

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The very wide range of values of CS obtained for CS for different groupings of paleotemperature data, for meteorological data, and for satellite data suggest either a widely varying feedback conditions for the different observational data sets, or the existence of additional factors beyond currently used feedbacks affecting global temperature.

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Table 1. Values of $\Delta\delta^{18}\text{O}$ and $\Delta p\text{CO}_2$ computed from data provided by Zachos et al. (1996) and Pearson et al. (2009).

Errors for $\Delta\delta^{18}\text{O}$ are computed from data points averaged.

Errors are not computed for Hole 522 since data is poor. Averages are indicative only.

Errors for $\Delta p\text{CO}_2$ are computed using confidence values provided for each data point by Pearson et al. (2009).

Data units: $\Delta\delta^{18}\text{O}$ is in ‰. $\Delta p\text{CO}_2$ is in ppmv.

HOLE 744				
Time segment		No. Of points	Average	error (1-sigma)
<i>a</i>		12	2.296	0.023
AVE(<i>a,c</i>)		26	2.314	0.018
<i>b</i>		11	2.147	0.024
AVE(<i>a,c</i>) – <i>b</i>	$\Delta\delta^{18}\text{O}$		0.166	0.023
	ΔT		0.594	0.063
HOLE 522 Cibicidoides				
Time segment		No. Of points	Average	
<i>a</i>		4	2.38	
<i>b</i>		7	2.27	
<i>a</i> – <i>b</i>	$\Delta\delta^{18}\text{O}$		0.1	
HOLE 522 Gyroidinoides				
Time segment		No. Of points	Average	
<i>a</i>		2	2.97	
<i>b</i>		5	2.81	
<i>a</i> – <i>b</i>	$\Delta\delta^{18}\text{O}$		0.15	
Kilwa Formation				
		No. Of points	Average	error (1-sigma)
$p\text{CO}_2$ baseline		2	774.5	53.6
$p\text{CO}_2$ high value		2	1132.5	86.6
	$\Delta p\text{CO}_2$		358	102

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Table 2. Some estimates of climate sensitivity published 2004–2012. Values are in units °C, for a doubling of $p\text{CO}_2$. Where published values are in units °C/(Wm⁻²), the published value is multiplied by 3.7 for the purpose of this comparison.

Year	Author	Source Data	Low	Median	High	Confidence
2004	Lea	0–400 Ka	4.4	5.2	6	95 %
2006	Forest et al.	climate data 90 and 40 yr	2.1	2.9	8.9	90 %
2007	IPCC AR4 (Solomon et al.)	multiple	2	3	4.5	66 %
2008	Chylek and Lohmann	LGM	1.3		2.3	95 %
2009	Douglas and Christy	Satellite obs, troposphere temp		1.1		
2010	Kohler et al.	Pleistocene glaciation	1.4	2.4	5.2	
2010	Pagani et al.	Pliocene paleotemperatures	7		9	
2011	Annan and Hargreaves	climate data 150 yr	2		4	95 %
2011	Lindzen and Choi	Satellite obs, sea surface temp	0.5	0.7	1.3	95 %
2011	Schmittner et al.	LGM	1.7	2.3	2.6	66 %
2012	Gillett	Climate data 160 yr	1.3		1.8	
2012	Rohling et al.	Pleistocene glaciation	1.7	3.1	5	66 %
2012	Lewis	climate data 90 and 40 yr	0.8	1.3	2.1	90 %
2012	Dowsett	Pliocene paleotemperatures	4		8	
2012	Hansen and Sato	LGM (fast feedbacks)	2	3	4	66 %
		LGM (incl slow millenium-scale feedbacks)	4	6	8	66 %
2012	Asten	Early Oligocene CO ₂ pulse	0.7	1.1	1.5	66 %

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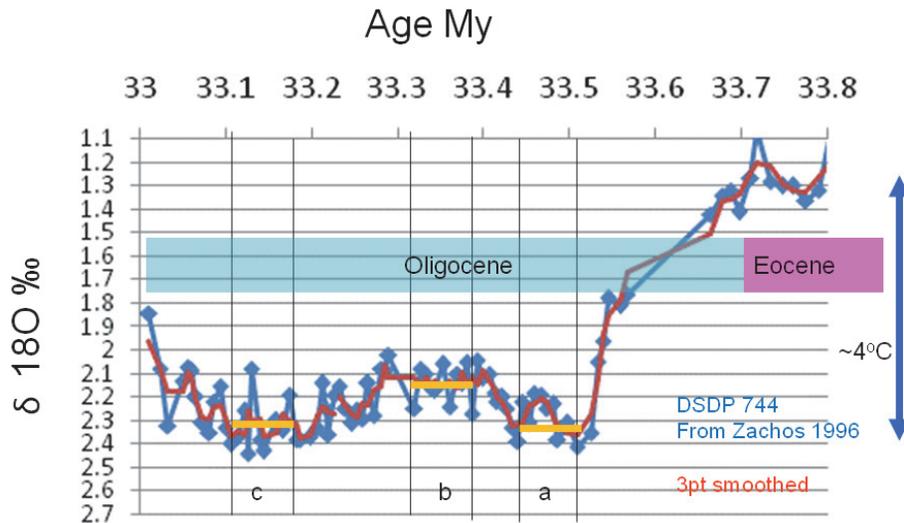


Fig. 1. Oxygen isotope data from Zachos et al. (1996) for DSDP hole 744 as original data (blue) and smoothed with a 3-point running mean (red). Vertical black bars indicate three data segments a,b,c, selected for base-line temperature estimates before, during, and after the CO₂ pulse shown in Fig. 3. Yellow lines show mean δ¹⁸O values for the selected segments.

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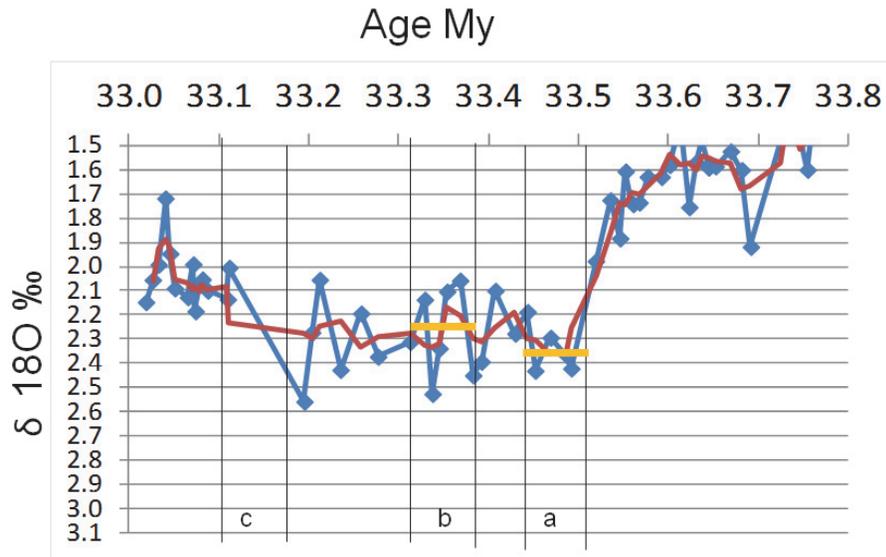


Fig. 2. Oxygen isotope data from Zachos et al. (1996) for DSDP hole 522, *Cibicidoides* species forams, as original data (blue) and smoothed with a 3-point running mean (red). Vertical black bars indicate the same data time segments as used in Fig. 1. Yellow lines show mean $\delta^{18}\text{O}$ values for the selected segments.

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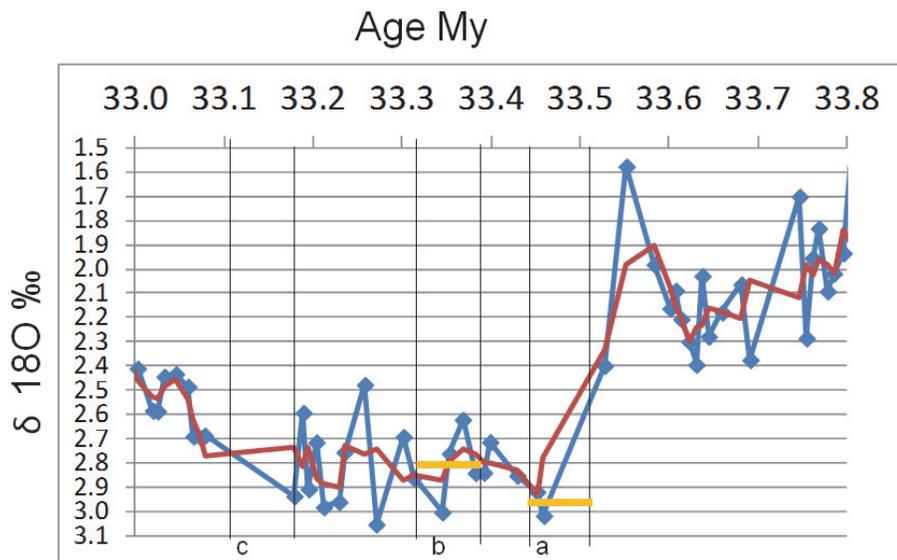


Fig. 3. Oxygen isotope data from Zachos et al. (1996) for DSDP hole 522, *Gyroidinoides* species forams, as original data (blue) and smoothed with a 3-point running mean (red). Vertical black bars indicate the same data time segments as used in Fig. 1. Yellow lines show mean $\delta^{18}\text{O}$ values for the selected segments.

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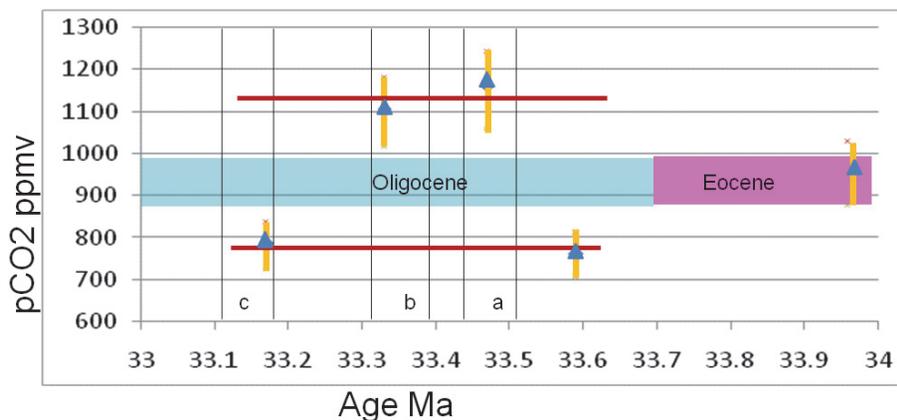


Fig. 4. Estimates of atmospheric $p\text{CO}_2$ from Pearson et al. (2009) from upper-ocean forams in the Kilwa formation, Tanzania. Vertical black bars indicate the same data time segments as used in Fig. 1. Yellow lines show 66 % confidence intervals from Pearson et al. Horizontal brown lines show average values for $p\text{CO}_2$ at the low-temperature limits, and for the maximum of the $p\text{CO}_2$ pulse.

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