

# The carbon cycle during the Mid Pleistocene Transition: the Southern Ocean Decoupling Hypothesis

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**Abstract.** Various hypotheses were proposed within recent years for the interpretation of the Mid Pleistocene Transition (MPT), which occurred during past 2 000 000 years (2 Myr). We here add to already existing theories on the MPT some data and model-based aspects focusing on the dynamics of the carbon cycle. We find that the average glacial/interglacial (G/IG) amplitudes in benthic  $\delta^{13}\text{C}$  derived from sediment cores in the deep Pacific ocean increased across the MPT by ~40%, while similar amplitudes in the global benthic  $\delta^{18}\text{O}$  stack LR04 increased by a factor of two over the same time interval. The global carbon cycle box model BICYCLE is used for the interpretation of these observed changes in the carbon cycle. Our simulation approach is based on regression analyses of various paleo-climatic proxies with the LR04 benthic  $\delta^{18}\text{O}$  stack over the last 740 kyr, which are then used to extrapolate changing climatic boundary conditions over the whole 2 Myr time window. The observed dynamics in benthic  $\delta^{13}\text{C}$  cannot be explained if similar relations between LR04 and the individual climate variables are assumed prior and after the MPT. According to our analysis a model-based reconstruction of G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  before the MPT is possible if we assume a different response to the applied forcings in the Southern Ocean prior and after the MPT. This behaviour is what we call the “Southern Ocean Decoupling Hypothesis”. This decoupling might potentially be caused by a different cryosphere/ocean interaction and thus changes in the deep and bottom water formation rates in the Southern Ocean before the MPT, however an understanding from first principles remains elusive. Our hypothesis is also proposing dynamics in atmospheric  $p\text{CO}_2$  over the past 2 Myr. Simulated  $p\text{CO}_2$  is vary-

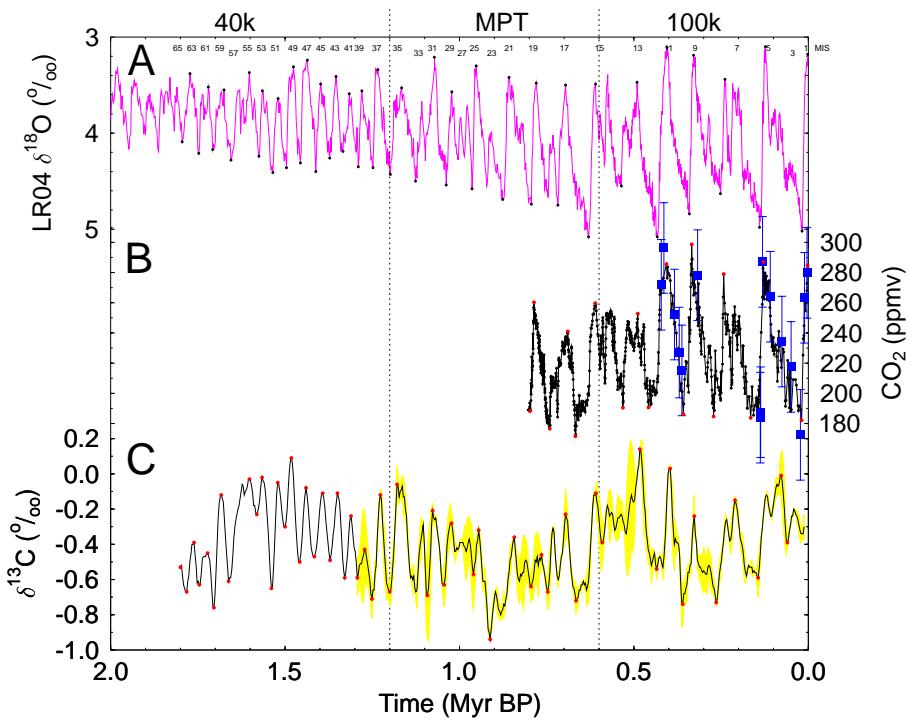
ing between 180 and 260  $\mu\text{atm}$  before the MPT. The consequence of our Southern Ocean Decoupling Hypothesis is that the slope in the relationship between Southern Ocean SST and atmospheric  $p\text{CO}_2$  is different before and after the MPT, something for which first indications already exist in the 800 kyr  $\text{CO}_2$  record from the EPICA Dome C ice core. We finally discuss how our findings are related to other hypotheses on the MPT.

## 1 Introduction

More than six decades ago Milutin Milankovitch proposed that variations in the orbital parameters of the Earth might be responsible for glacial/interglacial (G/IG) transitions in climate occurring on timescale of  $10^5$  to  $10^6$  years (Milankovitch, 1941). Thirty-five years later, the work of Hays et al. (1976) showed that similar frequencies of approximately 20 kyr, 40 kyr, and 100 kyr are found in the orbital variations and a deep ocean sediment record covering the last 430 kyr and thus Milankovitch’s idea was for the first time supported by a data set. Nevertheless, already in this first work which connected insolation and climate response, the power in the 1/100-kyr frequency of the orbital variations was much smaller than in the climate signal recorded in the sediment. Since then strong nonlinear feedbacks in the climate system are called for to explain this dominant 100-kyr frequency which is found in most climate records covering approximately the last 1 Myr (e.g. Imbrie et al., 1993). Furthermore, the origins of the 100-kyr cycles are interactions of different planets in our solar system leading to eccentricity anomalies in at least five different independent periods between 95 and 107 kyr (Berger et al., 2005). It is also long known that climate reconstructions which go further back in



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**Fig. 1.** Data evidences for changes in the climate system and the global carbon cycle over the last 2 Myr. **(A)** LR04 benthic  $\delta^{18}\text{O}$  stack (Lisiecki and Raymo, 2005). Odd MIS during the last 1.8 Myr are labelled. Those MIS (3, 23, 27, 33, 57) not used for the calculation of G/IG amplitudes due to its weak representation in both LR04 and the Pacific  $\delta^{13}\text{C}$  records are offset (similar as in Raymo et al., 2004). **(B)** Atmospheric CO<sub>2</sub>. Ice core measurements (circles, lines) from Vostok and EPICA Dome C (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008). Vostok CO<sub>2</sub> on the orbitally tuned age scale (Shackleton, 2000), EPICA Dome C on the EDC3\_gas.a age scale (scenario 4 in Louergue et al., 2007; Parrenin et al., 2007b). Reconstructed CO<sub>2</sub> (squares) based on pH in equatorial Atlantic waters derived from  $\delta^{11}\text{B}$  measured in planktic foraminifera (Hönisch and Hemming, 2005). **(C)** Mean benthic  $\delta^{13}\text{C}$  measured from ODP846 (Raymo et al., 2004) and ODP677 (Raymo et al., 1997). The data sets of both sediment cores were interpolated at 3 kyr steps, smoothed to a 3-point running mean and are synchronised on the orbitally tuned age scale of Shackleton et al. (1990). ODP677 covers only the last 1.3 Myr. Plotted here is the mean  $\delta^{13}\text{C} \pm 1 \text{ SD}$  (yellow). Black (A) and red (B, C) dots denote local minima/maxima within individual MIS, which were used to calculate G/IG amplitudes.

time do not show this 100-kyr variability but are dominated by the 40-kyr cycle caused by Earth's obliquity (Shackleton and Opdyke, 1976; Pisias and Moore Jr., 1981). Since then this shift in the climate from a 40-kyr variability in the Early Pleistocene (the 40 k world) towards a 100-kyr periodicity in the last several hundreds of thousands years (the 100 k world) was called the "Mid Pleistocene Transition (MPT)" and sometimes the "Mid Pleistocene Revolution". Besides this shift in the dominant frequency the MPT is also characterised by an increase in G/IG amplitudes in climate signals from the 40 k to the 100 k world, as clearly seen, for example, in the Lisiecki and Raymo (2005) LR04 benthic  $\delta^{18}\text{O}$  stack (Fig. 1A). A convincing theory which explains these observations remains elusive, however in recent years several hypotheses on the interpretation of the MPT were put forward (e.g. Maslin and Ridgwell, 2005; Raymo et al., 2006; Schulz and Zeebe, 2006; Clark et al., 2007; Huybers, 2007; Bintanja and van de Wal, 2008).

Interestingly, little attention has been given in most of these studies to changes in the carbon cycle. This might be based on the fact that ice core reconstructions including measurements of atmospheric CO<sub>2</sub> are so far restricted to the last 800 kyr covered in the Vostok and EPICA Dome C ice cores (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008). However, it has been shown that atmospheric CO<sub>2</sub> can be calculated from pH reconstructions based on boron isotopes (Hönisch and Hemming, 2005) and thus the limitation in the extension of the CO<sub>2</sub> time series given by the retrieval of old ice cores might at least be partially compensated in the near future (Fig. 1B). Furthermore, there is ample information on carbon cycle dynamics in published benthic  $\delta^{13}\text{C}$  reconstructions. From these benthic  $\delta^{13}\text{C}$  records at least long-term trends during the past 1.2 Myr were investigated recently (Hoogakker et al., 2006).

We here extend on the interpretation of carbon cycle dynamics across the MPT, but concentrate on the G/IG amplitudes. For this aim we perform simulations with the global carbon cycle box model BICYCLE. BICYCLE was the

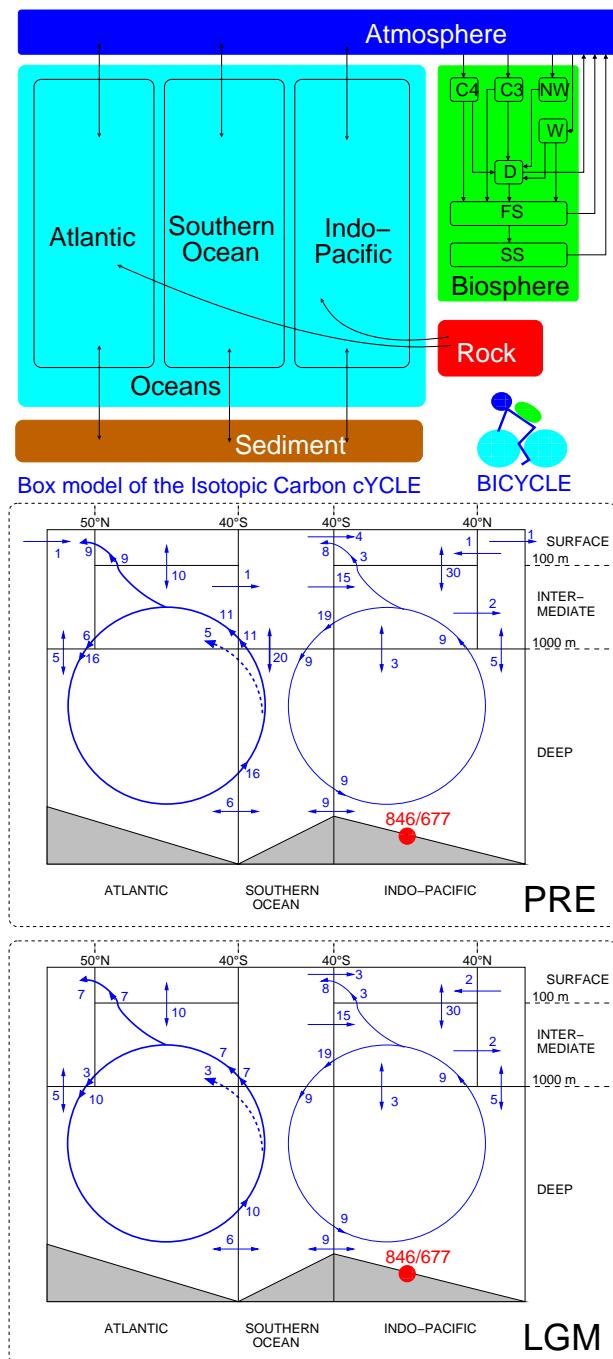
only full carbon cycle model used in the “EPICA challenge” (Wolff et al., 2004, 2005; Köhler and Fischer, 2006) “to predict based on current knowledge, what carbon dioxide” further back in time “will look like”. This challenge was started after the publication of climate signals of EPICA Dome C covering eight glacial cycles (EPICA-community-members, 2004), but before the presentation of any of the CO<sub>2</sub> data extending Vostok’s CO<sub>2</sub> record beyond 400 kyr BP (Siegenthaler et al., 2005; Lüthi et al., 2008). As older ice cores are planned to be drilled in the near future (Brook et al., 2006), this study can therefore be understood as an extension of the “EPICA challenge” further back in time. However, for the time being our results here will focus on the interpretation of measured benthic  $\delta^{13}\text{C}$  of the deep Pacific Ocean (Fig. 1C). Finally, we will discuss our results in the context of various recently published hypotheses on the causes of the MPT.

## 2 Methods

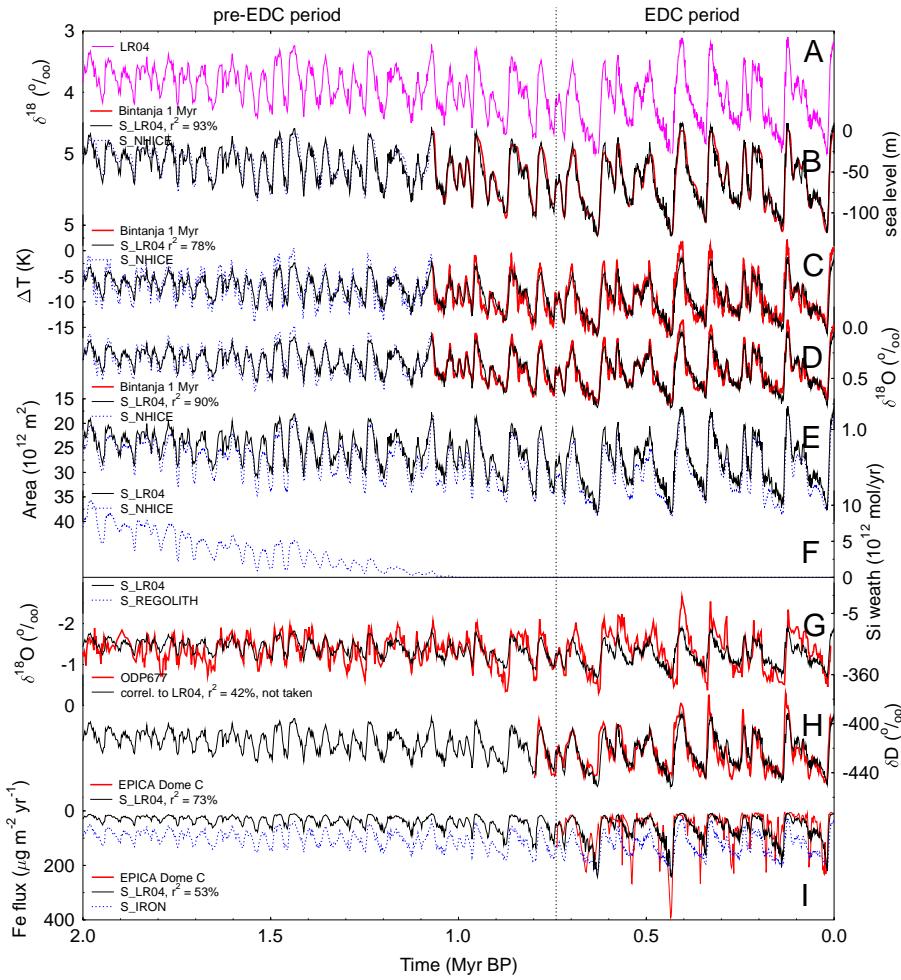
### 2.1 The model BICYCLE

To investigate the consequences of changes in climate on the carbon cycle across the MPT we use the carbon cycle box model BICYCLE (Fig. 2). It consists of a ten reservoir ocean module, one well mixed atmospheric box and a globally averaged terrestrial biosphere represented by seven boxes which distinguish C<sub>3</sub> and C<sub>4</sub> photosynthesis, and soils with different turnover times (Köhler and Fischer, 2004; Köhler et al., 2005). Prognostic variables are carbon (DIC in the ocean),  $\delta^{13}\text{C}$  and  $\Delta^{14}\text{C}$  in all boxes, and additionally alkalinity, PO<sub>4</sub> and O<sub>2</sub> in the ocean boxes. The model is based on former box models of the ocean (Munhoven, 1997) and the terrestrial biosphere (Emanuel et al., 1984), but was adapted and updated in previous studies. So far, BICYCLE was applied to understand carbon cycle dynamics during Termination I (Köhler and Fischer, 2004; Köhler et al., 2005), participated in the EPICA challenge (Wolff et al., 2005; Köhler and Fischer, 2006), and was used for the interpretation of atmospheric  $\delta^{13}\text{C}$  and  $\Delta^{14}\text{C}$  (Köhler et al., 2006a,b).

We use a model configuration which differs only slightly from previous applications in the definition of water fluxes between ocean reservoirs. In earlier applications all of the upwelling water in the Southern Ocean was travelling through the Southern Ocean surface box. Here, 30% of the upwelling flux is immediately relocated to the intermediate box in the equatorial Atlantic (dashed lines in Fig. 2). This is reasoned with the short residence time of these waters at the surface which is too short for equilibration with the atmosphere. With this model revision the results for atmospheric  $p\text{CO}_2$  are  $\sim 10 \mu\text{atm}$  lower during glacial maximums, which is still within the uncertainty range given by the ice core measurements (see Sect. 3.2). However, the revision brings the simulated  $\delta^{13}\text{C}$  in the deep ocean closer to paleo reconstructions.



**Fig. 2.** The BICYCLE carbon cycle model. Top: Main geometry. Middle and bottom: Geometry and ocean circulation fluxes (in  $\text{Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) of the oceanic module. PRE: preindustrial circulation based on the World Ocean Circulation Experiment WOCE (Ganachaud and Wunsch, 2000). LGM: assumed circulation during the LGM. The latitudinal position of the sediment cores in the equatorial Pacific ODP677/846 used for comparison is indicated by the red dot.

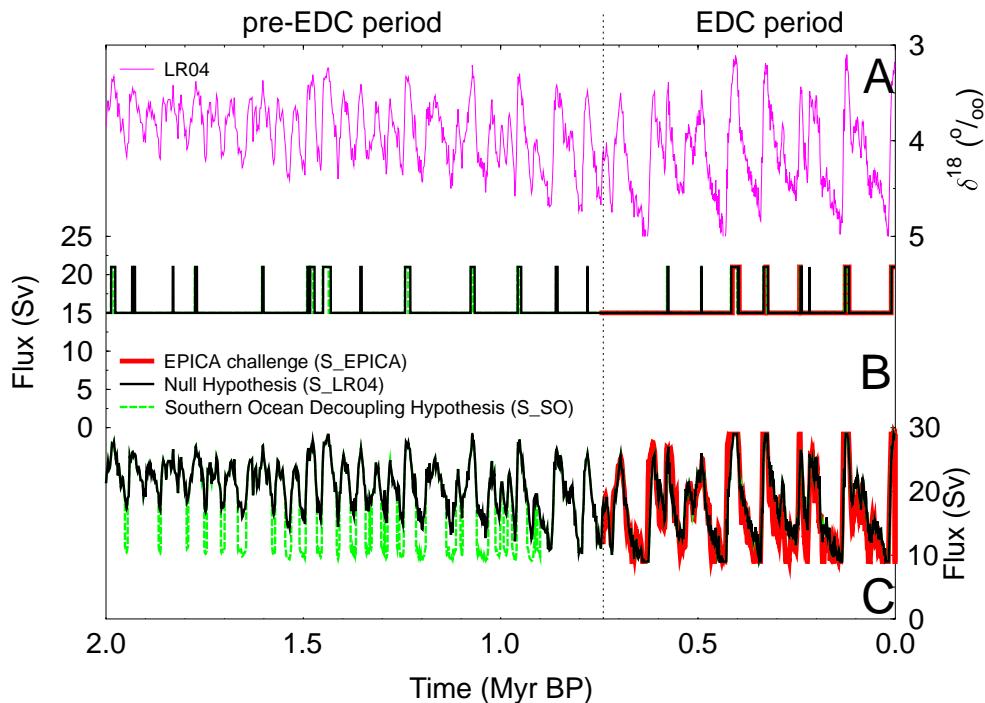


**Fig. 3.** Paleo-climatic records which were used to force the BICYCLE model. Original records (red, bold), those calculated from correlations with LR04, used in scenario S\_LR04 (black, thin), and alternative forcings (blue, broken). (A) LR04 benthic  $\delta^{18}\text{O}$  stack (Lisiecki and Raymo, 2005). (B) Sea level changes. (C) Temperature changes over land in the Northern Hemisphere ( $40\text{--}80^\circ\text{N}$ ). (D) Variability in deep ocean  $\delta^{18}\text{O}$  caused by deep ocean temperature changes. (E) Changes in areal extent of northern hemispheric ice sheets. (F) Additional silicate weathering flux due to regolith erosion. (G) Planktic  $\delta^{18}\text{O}$  of ODP677 ( $1^\circ 12'\text{N}, 83^\circ 44'\text{W}$ ) (Shackleton et al., 1990). (H) Sea level corrected deuterium  $\delta\text{D}$ . (I) Atmospheric iron fluxes to Antarctica. B–D: 1 Myr after Bintanja et al. (2005), 2–1 Myr BP and (E) after Bintanja and van de Wal (2008). H–I as measured in the EPICA Dome C ice core (EPICA-community-members, 2004; Wolff et al., 2006).

Assumed temporal changes in ocean circulation are (i) a highly stratified glacial Southern Ocean with less vertical exchange, (ii) a reduced North Atlantic Deep Water (NADW) formation and subsequent fluxes during glacials, and (iii) a closure of the Bering Strait during glacials caused by their sea level low stands (Fig. 2). The strength of the vertical mixing flux in the Southern Ocean is linearly coupled to the variability in the Southern Ocean SST (Fig. 3H). The mixing flux is not allowed to exceed preindustrial values and – to be consistent with previous studies (Köhler et al., 2005) – is not allowed below its assumed minima at LGM (Fig. 4C). NADW formation and Bering Strait outflow switch only between glacial and interglacial states. Their changes are trig-

gered by northern hemispheric temperature (Fig. 3C). Thus, the strong Atlantic overturning depicted in Fig. 2A exists only during peak interglacial conditions (Fig. 4B).

For the analysis of the  $^{13}\text{C}$  cycle in the deep ocean a sediment box with initially 50 000 PgC and a  $\delta^{13}\text{C}$  of  $2.75\text{\textperthousand}$  is introduced in each deep ocean basin (Atlantic, Southern Ocean, Indo-Pacific). The initial  $\delta^{13}\text{C}$  of the sediments is similar to the long-term average signature of the  $\text{CaCO}_3$  produced in the surface ocean. Initial values were chosen such that the  $\delta^{13}\text{C}$  signature of the sediments does not present any large drift, they change in the chosen setting less than  $\pm 0.04\text{\textperthousand}$  over the simulation period. If the initial  $\delta^{13}\text{C}$  in the sediments is, for example, smaller than that of the exported



**Fig. 4.** Time series of surface to deep ocean fluxes in the North Atlantic (**B**) and the Southern Ocean (**C**) for different scenarios. Note, that plotted numbers sum up all fluxes, e.g. deep water formation and vertical mixing fluxes. After 0.9 Myr BP the fluxes in the Southern Ocean for S\_LR04 and S\_SO are identical. (**A**) LR04 benthic  $\delta^{18}\text{O}$  stack (Lisiecki and Raymo, 2005) for comparison.

and accumulated  $\text{CaCO}_3$ , sedimentary  $\delta^{13}\text{C}$  increases over time (e.g. +0.25‰ over 2 Myr for initially 1‰ lower  $\delta^{13}\text{C}$  in the sediments), while that of the ocean/atmosphere/biosphere decreases (mean ocean by –0.02‰ for the same example). The  $\text{CaCO}_3$  which enters the deep ocean boxes and can be accumulated in the sediments, was produced with a fixed ratio between export of organic matter (OM) and hard shells in the surface waters ( $\text{C}_{\text{OM}}:\text{C}_{\text{CaCO}_3}=10:1$ ). OM export at 100 m depth is prescribed for present day to 10 PgC (e.g. Schlitzer, 2000), but depends on available macro-nutrients. This is realised with maximum productivity in equatorial waters. Increased export might occur in the Southern Ocean, if macro-nutrients are available and the proxy for iron input into the Southern Ocean suggests the stimulation of additional productivity due to iron fertilisation. The chosen export production of organic matter and rain ratio lead to modern  $\text{CaCO}_3$  export of 1 PgC  $\text{yr}^{-1}$  similar to other applications (e.g. Jin et al., 2006).  $\text{CaCO}_3$  is partially (20%) remineralised in the intermediate layers. In the deep ocean carbonate compensation is calculated with a relaxation approach resulting in either sedimentation of  $\text{CaCO}_3$  or dissolution of sediments as a function of the offset from initial (present day)  $\text{CO}_3^{2-}$  concentrations. This relaxation approach considers a slow time delayed response of the sediments to deep ocean carbonate ion anomalies with an e-folding time of 1.5 kyr to be in line with data-based reconstructions (Marchitto et al., 2005).

During sedimentation and dissolution no isotopic fractionation is assumed:  $\text{CaCO}_3$  which is added to the sediments has the same  $\delta^{13}\text{C}$  as during hard shell production at the water surface, while the dissolved carbonate carries the  $\delta^{13}\text{C}$  signal of the sediment box.

## 2.2 Time-dependent forcing

For the applications of BICYCLE over the last 740 kyr (Köhler and Fischer, 2006) various proxy data sets derived from sediment and ice cores were used to force it with time-dependent climatic boundary conditions. The success of this approach depends heavily on the synchronisation of the used data sets onto a common time scale. Due to this intrinsic feature of the forcing mechanisms and the restriction of Antarctic ice core records to the last 800 kyr we relied on a simple, but more consistent approach to force our model over the last 2 Myr.

This new approach is based on the use of the benthic  $\delta^{18}\text{O}$  stack LR04 derived by Lisiecki and Raymo (2005) as a master record for any observed climatic change. We calculate regression functions between LR04 and all those records used to force the model over the last 740 kyr (Table 1 and Figs. A1–A6 in the Supplemental Material <http://www.clim-past.net/4/311/2008/cp-4-311-2008-supplement.pdf>). These regression

**Table 1.** Regression functions calculated between LR04 benthic  $\delta^{18}\text{O}$  (variable  $x$ ) (Lisiecki and Raymo, 2005) and the mentioned paleo records. All EPICA Dome C records were taken on the EDC3 age scale (Parrenin et al., 2007a). Complete time series are found in the supplement.

Record (variable $y$ )	Symbol	Length (kyr)	Regression function	$r^2$ (%)	Reference
Sea level	LR04-SEAL	1070	$y=234.51-71.41 \cdot x$	93	Bintanja et al. (2005)
Northern hemispheric temperature	LR04-NHdT	1070	$y=22.74-7.75 \cdot x$	78	Bintanja et al. (2005)
$\Delta\delta^{18}\text{O}$ of LR04 caused by deep sea temperature	LR04-DEEPdT	1070	$y=-1.12-0.38 \cdot x$	90	Bintanja et al. (2005)
$\delta\text{D}$ in EPICA Dome C (SO SST proxy)	LR04-dD	800	$y=-298.00-30.10 \cdot x$	73	Jouzel et al. (2007)
Fe flux in EPICA Dome C (SO Fe fertilisation proxy)	LR04-FE1	740	$y=10^{-1.36+0.74} \cdot x$	53	Wolff et al. (2006)
Fe flux in EPICA Dome C (SO Fe fertilisation proxy) <sup>a</sup>	LR04-FE2	740	$y=-279.34+98.12 \cdot x$	13	Wolff et al. (2006)
Planktic $\delta^{18}\text{O}$ in ODP667 (equatorial SST proxy)	LR04-EQSST	2000	$y=-3.84-0.63 \cdot x$	42	Shackleton et al. (1990)

<sup>a</sup> Linear regression only for points with Fe flux  $\geq 100 \mu\text{g m}^{-2} \text{yr}^{-1}$ .

functions are then used to extrapolate how the various components of the Earth's climate, which are used as changing boundary conditions in our carbon cycle model, might have changed over the last 2 Myr. Although this approach neglects any existing leads and lags between various parts of the climate system, it should in the light of the relatively coarse temporal resolution of LR04 ( $\Delta t=1-2.5$  kyr between 2 Myr and present) be a good approximation to estimate G/IG changes. Furthermore, it implies that the correlation of the various paleo records with LR04 were in principle not different before and after the MPT and that the undertaken extrapolation of the forcings to 2 Myr is meaningful. It also implies that climate processes and their impacts on the carbon cycle were following similar functional dependencies in the 40 k and the 100 k world. Therefore, we refer to this approach, which is heavily based on forcing functions derived from LR04 (scenario S\_LR04) as our "Null Hypothesis". The comparison of model results of this Null Hypothesis (S\_LR04) with those of the scenarios forced with the original records (S\_EPICA: original 740 kyr application; S\_EPICA+: 740 kyr application with revised ocean circulation) gives us evidences how much variability in the carbon cycle will be lost by the simplification of the forcing mechanisms. Thus, this comparison represents a sort of "ground-truthing" which is important for the interpretation of the results going further back in time than 740 kyr BP.

The correlation of the regressions functions between LR04 and the other records is in general high ( $r^2$  of 76% to 93%, Table 1). The poorest correlations ( $r^2$  of 42% and 52%) exist for LR04 and planktonic  $\delta^{18}\text{O}$  in ODP667 (a proxy for equatorial SST) and for LR04 and the iron flux as measured in the EPICA Dome C ice core (a proxy for iron input into the Southern Ocean). Due to this relatively bad correlation and the fact that the record of planktonic  $\delta^{18}\text{O}$  in ODP667 is available over the whole 2 Myr time window we refrain from using the LR04-based substitute, but use the original  $\delta^{18}\text{O}$  data from ODP667 throughout our simulations. Fur-

thermore, it was shown (Liu et al., 2008) that tropical SST dynamics across the MPT are different than global climate variations contained in LR04. This is another argument to rely on the original record in the equatorial region.

The second correlation with low  $r^2$  determines variations in the size of the marine export production in the Southern Ocean and is responsible for the simulated rise of atmospheric  $p\text{CO}_2$  of up to  $20 \mu\text{atm}$  during Termination I (Köhler et al., 2005). Furthermore, for all forcing records but the iron flux record linear regression functions led to adequate result. The regression between the iron flux and LR04 needs the use of an exponential regression function. Because of the poor correlation of these two records and the potential consequences of this for biologically driven carbon export to the ocean interior an alternative scenario S\_IRON is applied, in which changes during peak iron fluxes are better represented than previously. This is achieved through a linear regression between LR04 and the iron flux that is restricted to iron fluxes  $\geq 100 \mu\text{g m}^{-2} \text{yr}^{-1}$  (Fig. 3I). These iron peaks are of special interest because it can be assumed that during these times the Southern Ocean marine biology was not limited by iron and thus export production was enhanced (Martin, 1990; Parekh et al., 2008).

Furthermore, changes in various climate variables (sea level, northern hemispheric temperature, deep sea temperature) were already estimated out of LR04 using an inverse modelling approach (Bintanja et al., 2005) for the 740 kyr long application of the BICYCLE model. The basis of their approach is the deconvolution of the temperature and sea level information contained in  $\delta^{18}\text{O}$  of the LR04 stack. Bintanja and van de Wal (2008) use the same methodology, but extend the analysis 3 Myr back in time. Here, the North American ice sheets play a central role in intensifying and prolonging the glacial cycles during the MPT. Long-term climate cooling enables the North American ice sheets to grow in the 100 k world to a stage in which they are able to merge, after which they can grow even more rapidly until basal-

**Table 2.** Summary and description of simulation scenarios.

Name	Forcing	Comment
740 kyr simulations (100 k world only)		
S_EPICA	as in original EPICA challenge, forced with different paleo records	published in Köhler and Fischer (2006)
S_EPICA+	similar to S_EPICA, but with revised Southern Ocean upwelling (Fig. 2)	
Scenarios across MPT based on previous hypotheses		
S_LR04	forced with time series derived via correlation and regression function between LR04 and climate records used in S_EPICA	Our <i>Null Hypothesis</i>
S_IRON	as S_LR04, but with alternative regression function for the input of iron in the Southern Ocean with consequences for marine export production	
S_NHICE	as S_LR04, but the smaller G/IG amplitudes in sea level change in the 40 k world are mainly caused by Northern Hemisphere ice volume anomaly. Sea level, deep ocean and northern hemispheric temperature, and areal extent of ice sheets are not obtained from correlation with LR04, but taken from the cited simulation study	input from Bintanja and van de Wal (2008)
S_REGOLITH	as S_NHICE, but additionally changing silicate weathering rates between 2 and 1 Myr BP are considered	following the <i>Regolith Hypothesis</i> of Clark et al. (2007)
S_COM	combining S_LR04 with improvements of scenarios S_IRON, S_NHICE, and S_REGOLITH	
Scenarios across MPT based on our new Southern Ocean Decoupling Hypothesis		
S_SO	as S_LR04, but with revised (larger than in S_LR04) G/IG amplitudes in the Southern Ocean vertical mixing rates before 900 kyr BP	Our <i>Southern Ocean Decoupling Hypothesis</i>
S_FINAL	combining S_COM with S_SO, or improving the <i>Null Hypothesis</i> with the alternative regression function for iron input in the Southern Ocean (S_IRON), details on northern hemispheric ice sheet evolution (S_NHICE), the <i>Regolith Hypothesis</i> (S_REGOLITH) and the <i>Southern Ocean Decoupling Hypothesis</i> (S_SO)	our final (best guess) scenario

sliding related instabilities in this huge ice sheet causes catastrophic collapse and deglaciation. The variables mentioned above and changes in northern hemispheric ice sheet area, necessary for the areal extent of the terrestrial biosphere, calculated by Bintanja and van de Wal (2008) out of LR04, will be used alternatively in scenario S\_NHICE (Fig. 3B–E).

Further changes in the carbon cycle are performed in scenario S\_REGOLITH. Following the *Regolith Hypothesis* of Clark et al. (2007) we assume that the regolith layer located beneath the northern hemispheric ice sheets got eroded over time before the MPT. This leads to an additional flux of silicate weathering or  $\text{HCO}_3^-$  input to the ocean. We assume a weathering and thus  $\text{HCO}_3^-$  flux, which declines over time (from  $12 \times 10^{12} \text{ mol C yr}^{-1}$  (2 Myr BP) to  $0 \text{ mol C yr}^{-1}$  – 1 Myr BP), and which is modulated by the areal extent of the northern hemispheric ice sheets (Fig. 3F). The carbonate chemistry of the ocean (including the magnitude of the carbonate compensation) is effected by these fluxes as they change the overall budgets of alkalinity and DIC. These numbers consider only the additional changes in the weathering rate, thus background silicate and carbonate weathering is implicitly included in our carbonate compensation mecha-

nism. It has been shown that to obtain stable atmospheric  $\text{CO}_2$  on time scales longer than glacial cycles volcanic outgassing and weathering fluxes balance each other (e.g. Zeebe and Caldeira, 2008). This implies that half of the carbon consumed by silicate weathering need to be supplied by volcanic out-gassing of  $\text{CO}_2$ , while the other half is taken from atmospheric  $\text{CO}_2$  (Munhoven and François, 1996, and references therein). The strength of the assumed fluxes are somewhat different than those used in Clark et al. (2007). The magnitude of the additional  $\text{HCO}_3^-$  input at 2 Myr BP ( $12 \times 10^{12} \text{ mol C yr}^{-1}$ ) was chosen to be of similar size as the estimated present-day fluxes given by Gaillardet et al. (1999). This would imply that at 2 Myr BP, at the time of maximum input of silicate weathering from regolith erosion, the amplitude of this process is twice that during present day. According to our understanding this would be a rough conservative estimate of the upper end of what impact can be expected from the *Regolith Hypothesis* on the global carbon cycle.

For the time being these scenarios, which are all well supported by other studies, were chosen as starting point. We will in the evaluation of these scenarios (Sects. 3.3–3.4) have to conclude that they are insufficient to describe the actual

**Table 3.** A compilation how boundary conditions (climate variables) were substituted in this application.

Climate variable	Substitution used in S_LR04 <sup>a</sup>	Alternative formulation	Scenarios which use alternative formulation	Fig. <sup>b</sup>
Sea level	LR04-SEAL	2 Myr simulation results <sup>c</sup>	S.NHICE, S.REGOLITH, S.COM, S.FINAL	B
SST (North Atlantic) <sup>d</sup>	$f(\text{LR04-NHDT})$	2 Myr simulation results <sup>c</sup>	S.NHICE, S.REGOLITH, S.COM, S.FINAL	C
SST (Equatorial Atlantic)	no substitution, take original record (ODP677)			G
SST (Southern Ocean) <sup>e</sup>	$f(\text{LR04-dD})$	Mixing decoupled from SO SST <sup>e</sup>	S.SO, S.FINAL	H
SST (Equatorial Pacific)	no substitution, take original record (ODP677)			G
SST (North Pacific)	$f(\text{LR04-NHDT})$	2 Myr simulation results <sup>c</sup>	S.NHICE, S.REGOLITH, S.COM, S.FINAL	C
Intermediate and deep ocean T	LR04-DEEPdT	2 Myr simulation results <sup>c</sup>	S.NHICE, S.REGOLITH, S.COM, S.FINAL	D
Fe fertilisation Southern Ocean	LR04-FE1	LR04-FE2	S.IRON, S.COM, S.FINAL	I
Northern hemispheric temperature	LR04-NHdT	2 Myr simulation results <sup>c</sup>	S.NHICE, S.REGOLITH, S.COM, S.FINAL	C
Northern hemispheric ice sheet area	$f(\text{LR04-SEAL})$	2 Myr simulation results <sup>c</sup>	S.NHICE, S.REGOLITH, S.COM, S.FINAL	E
Silicate weathering (regolith erosion)	–	linear decline × ice sheet area	S.REGOLITH, S.COM, S.FINAL	F

<sup>a</sup> Symbols taken from Table 1.

<sup>b</sup> Notation of sub-figures of Fig. 3 where the relevant time series are plotted.

<sup>c</sup> Bintanja and van de Wal (2008).

<sup>d</sup> North Atlantic deep water formation is coupled to North Atlantic SST.

<sup>e</sup> Southern Ocean vertical mixing is coupled to SST. This coupling is linear for 900–0 kyr BP, but follows at different pattern in earlier times. Before 900 kyr BP the reduction in vertical mixing in the Southern Ocean in glacials is larger than in the SST substitute. Vertical mixing (Fig. 4C) then follows a synthetic time series, which was derived in four steps:

(1) The surrogate of LR04 for Southern Ocean SST  $f(\text{LR04-dD})=y_1$  is linearly transformed to [0, 1]:

$y_2=f(y_1)$  with  $y_2(t=0 \text{ kyr BP})=0$  and  $y_2(t=18 \text{ kyr BP})=1$ .

(2) Stretch  $y_2$  to colder climates by  $y_3=(y_2)^3$ . This shifts, for example, the mid point  $y_2=0.5$  to  $y_3=0.125$ .

(3) Transfer stretched record linearly back to the range of SST values:  $y_4=f(y_3)$ .

(4) Restrict variations in vertical mixing to the range given by SST found at present and 17 kyr BP, as deduced for Termination I in Köhler et al. (2005).

variability in benthic  $\delta^{13}\text{C}$ , and at the end we will therefore make suggestions to further modify our assumptions and propose an new explanation, the *Southern Ocean Decoupling Hypothesis* (Sect. 3.5). All scenarios are summarised in Table 2. Details on all correlations and which variables were substituted in each scenario are described in the Tables 1 and 3 and in the Supplemental Material <http://www.clim-past.net/4/311/2008/cp-4-311-2008-supplement.pdf>. A detailed description how changing climatic boundary conditions impact on our carbon cycle model is published in Köhler and Fischer (2006).

### 3 Results

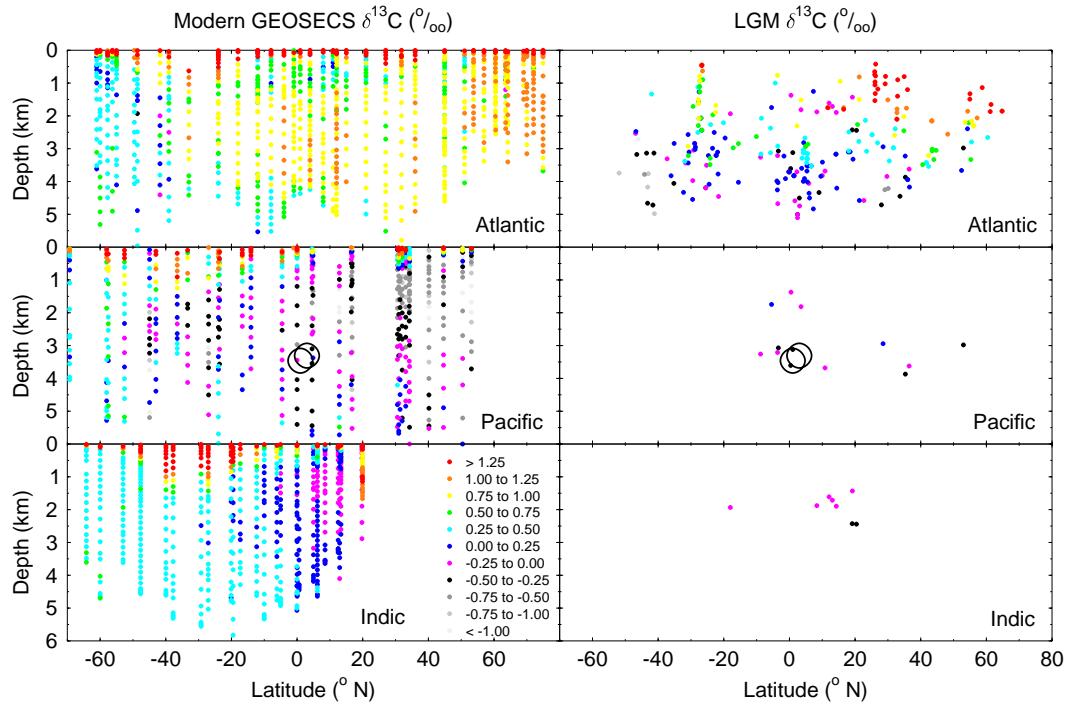
#### 3.1 Evidences from paleo records

We concentrate our paleo data analysis across the MPT on

1. the LR04 benthic  $\delta^{18}\text{O}$  stack (Lisiecki and Raymo, 2005) as global recorder of climate change (Fig. 1A),
2. atmospheric  $\text{CO}_2$  measured in ice cores (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008) (Fig. 1B),
3. reconstructed  $\delta^{13}\text{C}$  from the deep Pacific (Fig. 1C). Here, the deep Pacific  $\delta^{13}\text{C}$  is represented by an average

of benthic  $\delta^{13}\text{C}$  measured in two cores from the equatorial Pacific (ODP846: 3° S, 91° W, 3307 m water depth, Raymo et al., 2004; ODP677: 1° S, 83° W, 3461 m water depth, Raymo et al., 1997). They are plotted on an orbital tuned age scale (Shackleton et al., 1990), were interpolated to a uniform 3 kyr spacing and smoothed with a 3-points running mean, as performed already in Raymo et al. (2004).

We focus only on changes in  $\delta^{13}\text{C}$  in the deep Pacific Ocean, because changes in the Atlantic might depend largely on the core site due to changing deep and bottom water fluxes between glacial and interglacial times (Kroopnick, 1985; Curry and Oppo, 2005). These detailed changes in ocean circulation and the consequences for local  $\delta^{13}\text{C}$  can not be represented in our model due to the coarse spatial resolution. For the Southern Ocean not enough data sets exist to compile one record, which would be a representative of the whole Southern Ocean as it is defined in our model (south of 40° S). Those long records of which we are aware are all located around 40° S from the South Atlantic/Atlantic sector of the Southern Ocean. In the deep eastern Pacific, where our Pacific sites are located, the horizontal and vertical gradients in  $\delta^{13}\text{C}$  in the modern ocean are very small (Kroopnick, 1985) (Fig. 5). A similar uniform distribution of  $\delta^{13}\text{C}$  exists in large parts of the glacial Pacific (Boyle, 1992) (Fig. 5). Therefore, observed changes are assumed to be representative of basin



**Fig. 5.** Modern  $\delta^{13}\text{C}$  data and LGM  $\delta^{13}\text{C}$  reconstructions as function of ocean basin, depth and latitude. Left: GEOSECS data ( $\delta^{13}\text{C}$  of DIC in the water column, Kroopnick (1985)). Right: LGM reconstructions ( $\delta^{13}\text{C}$  of benthic foraminifera in sediment cores, Boyle, 1992; Bickert and Mackensen, 2004; Curry and Oppo, 2005). The reconstructions of Bickert and Mackensen (2004) are corrected for the phytodetritus effect. Circles mark the site locations of ODP846 and ODP677.

wide variations and not merely a recorder of local changes in ocean circulation. Purely local effects should be minimised by averaging two different cores.

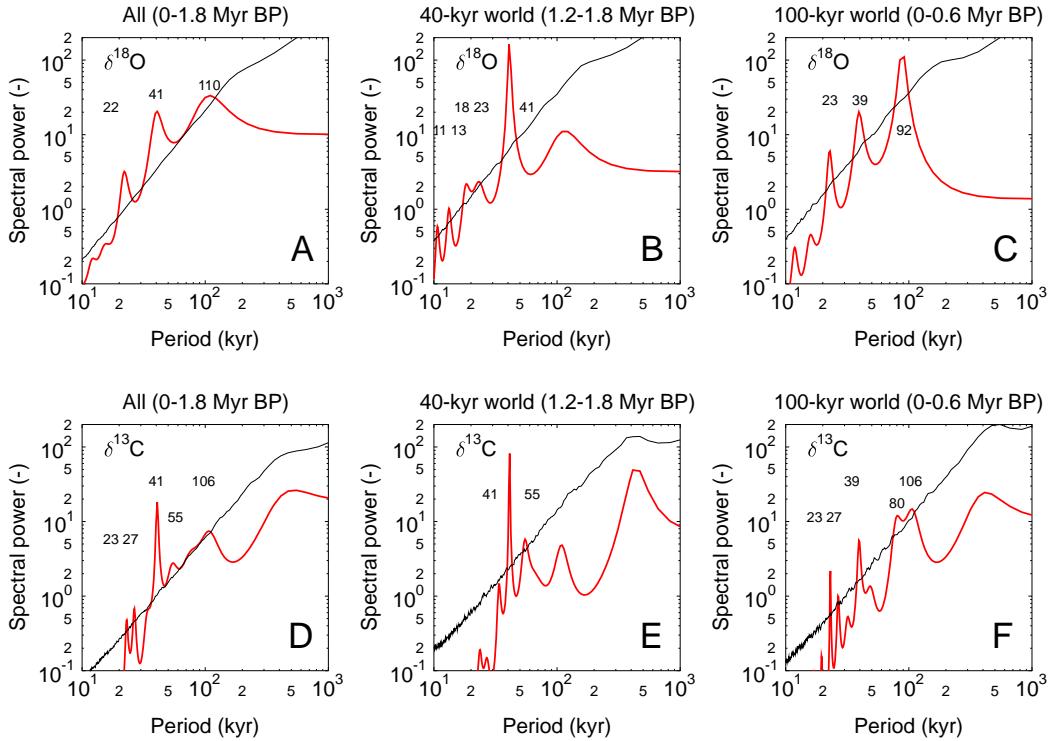
For data analysis and the following data-model comparison we divide our time period of interest in three time windows: (a) the 40 k world (1.8 to 1.2 Myr BP), (b) the MPT (1.2 to 0.6 Myr BP), and (c) the 100 k world (after 0.6 Myr BP). These are the same intervals as in Raymo et al. (2004) to allow comparison. Simulation results between 2.0 and 1.8 Myr BP are omitted in our further analysis due to the missing benthic  $\delta^{13}\text{C}$  data.

We use the maximum entropy spectral analysis (MESA) (Ghil et al., 2002) to clearly identify the MPT with its shift from 40-kyr to 100-kyr periodicity in the LR04  $\delta^{18}\text{O}$  stack (Fig. 6A–C, see also Lisiecki and Raymo, 2007). The benthic  $\delta^{13}\text{C}$  in the deep Pacific as representative of the carbon cycle does also record this transition from the 40 k to the 100 k world (Fig. 6D–F). There is also an even slower variability with a frequency of  $\sim 1/500 \text{ kyr}^{-1}$  superimposed, however this frequency component is not statistically significant within our MESA approach. Nevertheless, a low frequency component of  $1/400 \text{ kyr}^{-1}$  in  $\delta^{13}\text{C}$  of the deep Pacific was already identified over the last 2.4 Myr as one of the most important frequencies in ODP677 and ODP849 ( $0^\circ \text{ N}$ ,

$110^\circ \text{ W}$ , 3851 m water depth) (Mix et al., 1995). Furthermore, a long-term cyclicity of  $\sim 500 \text{ kyr}$  in  $\delta^{13}\text{C}$  has been found in all ocean basins during the Pleistocene (Wang et al., 2004). Spectral analysis of atmospheric  $\text{CO}_2$  is omitted due to data-limitation.

The sizes of the G/IG amplitudes of the selected records for the different time windows are of special interest in this study. They are summarised together with simulation results in Table 4. Besides the averages ( $\pm$ one standard deviation) their relative sizes during earlier times with respect to the 100 k world is investigated. This information is expressed in the so-called  $f$ -ratio ( $f_X = \frac{\bar{\Delta}x}{\Delta_{100k}} \cdot 100$  in %, with  $X = \text{MPT}$  or  $40 \text{ k}$ ). It gives information on the changes in amplitude over the MPT, and not in frequency. It will be used widely in the following to compare the qualitative behaviour of our simulations with the reconstructions. We have to acknowledge that for this analysis of G/IG amplitudes some periods (MIS 23, 27, 33, and 57), in which no distinct maxima could be identified in the  $\delta^{13}\text{C}$  records were omitted for further analysis. All local minima and maxima used here are marked in Fig. 1.

The global climate as represented by LR04 exhibits G/IG amplitudes, which increase by up to a factor of two over the MPT. In other words, the  $f$ -ratios are 76% and 51% for the



**Fig. 6.** Maximum entropy spectral analysis (MESA). Power spectra of the LR04 benthic  $\delta^{18}\text{O}$  stack (**A–C**) and the deep Pacific  $\delta^{13}\text{C}$  reconstruction (**D–F**) for different time windows, (time series plotted in Fig. 1A, C). Data (bold red) and their 99% confidence (thin black). Significant periods of labelled.

MPT and the 40 k world, respectively. Similarly, the G/IG amplitudes in the stacked Pacific  $\delta^{13}\text{C}$  increase over time, but not as much as the climate signal seen in LR04. Its amplitude increases from  $0.40 \pm 0.16\%$  (40 k) via  $0.44 \pm 0.15\%$  (MPT) to  $0.55 \pm 0.03\%$  (100 k), corresponding to  $f$ -ratios of 72% (40 k) and 80% (MPT), respectively. The same analysis is performed for the two individual ODP  $\delta^{13}\text{C}$  time series which were averaged here, to check if and how the stacking of both  $\delta^{13}\text{C}$  records leads to changes in the G/IG amplitudes. Indeed, the amplitudes in ODP677 in the 100 k world are with  $0.69 \pm 0.09\%$  larger than those of ODP846 ( $0.49 \pm 0.12\%$ ). They seemed to be of similar amplitude in the 40 k world ( $0.39\%$  vs.  $0.41\%$ ), but ODP677 covers only two G/IG transitions here. The consequence is that the  $f_{40\text{k}}$ -ratios differ (56% and 83% for ODP677 and ODP846, respectively). However, because of the shortness of ODP677 (1.3 Myr), we think the statement that Pacific benthic  $\delta^{13}\text{C}$  changed less in G/IG amplitudes across the MPT than LR04 is based on solid evidences.

Our data-based knowledge on variations in atmospheric  $\text{CO}_2$  is limited (Fig. 1B). Direct measurements of  $\text{CO}_2$  on air enclosures in ice cores is restricted to the last 800 kyr (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008). Within this time  $\text{CO}_2$  varies between 170 and 300 ppmv (the partial pressure of  $p\text{CO}_2$  of 170 to 300  $\mu\text{atm}$ ). Fur-

ther evidences on  $\text{CO}_2$  variability before the MPT do not exist. One alternative approach of reconstructing  $\text{CO}_2$  is based on the surface seawater pH, which itself is calculated out of  $\delta^{11}\text{B}$  measured on planktic foraminifera (Hönisch and Hemming, 2005). Existing  $\text{CO}_2$  reconstructions are consistent with ice core measurements, but with a large error of  $\pm 30 \mu\text{atm}$  (Fig. 1B). This approach has nevertheless the potential to extend the  $\text{CO}_2$  ice core records further back in time in the near future. For the time being we restrict our analysis of G/IG amplitudes in atmospheric  $\text{CO}_2$  to the ice core records. In the 100 k world the mean G/IG amplitude in  $\text{CO}_2$  is  $95 \pm 16$  ppmv. This is reduced to  $75 \pm 10$  ppmv in the MPT ( $f_{\text{MPT}}=79\%$ ), but we have to be aware that the ice cores contain only three G/IG transitions in the MPT time window (Table 4).

### 3.2 Ground-truthing of our approach based on LR04 – simulations for the last 740 kyr

Detailed discussions of simulation results obtained with BICYCLE over the last 740 kyr were already published. The previous application concentrated on atmospheric  $p\text{CO}_2$  (scenario S\_EPICA, Köhler and Fischer, 2006). The ocean circulation field used in the model was revised between the earlier and the present application for an enhanced represen-

**Table 4.** Analysis of G/IG amplitudes in LR04,  $p\text{CO}_2$ , and deep Pacific  $\delta^{13}\text{C}$  for reconstructions and simulation results divided into different time windows. The  $f$ -ratio ( $f_X = \frac{\bar{\Delta}X}{\bar{\Delta}_{100k}} \cdot 100$  with  $X = \text{MPT}$  or  $40\text{k}$ ) describes the relative size (in %) of G/IG amplitudes in comparison to the  $100\text{k}$  world. The number  $n$  of G/IG transitions included in these calculations are 6 ( $100\text{k}$ ), 8 ( $\text{MPT}$ ), and 13 ( $40\text{k}$ ) in all entries. Only in the  $\delta^{13}\text{C}$  sediment cores  $n$  is smaller: ODP677/846 has  $n=5$  ( $100\text{k}$ ) due to missing data during the Holocene and thus no G/IG values for Termination I, and in ODP677  $n=2$  ( $40\text{k}$ ). Most important scenarios (our Null Hypothesis S\_LR04 and our best guess scenario S\_FINAL) are highlighted in bold.

Name	100 k		MPT		40 k	
	$\bar{\Delta} \pm \text{SD}$	$\bar{\Delta} \pm \text{SD}$	$f_{\text{MPT}}$	$\bar{\Delta} \pm \text{SD}$	$f_{40\text{k}}$	
<b>global climate (‰)</b>						
LR04	$1.60 \pm 0.34$	$1.22 \pm 0.20$	76	$0.82 \pm 0.15$	51	
<b>atmospheric <math>p\text{CO}_2</math> (μatm)</b>						
Vostok + EDC	$95 \pm 16$	$75 \pm 10^a$	79	–	–	
S_EPICA	$70 \pm 11$	– <sup>b</sup>	–	–	–	
S_EPICA+	$81 \pm 11$	– <sup>b</sup>	–	–	–	
<b>S_LR04</b>	<b><math>70 \pm 18</math></b>	<b><math>51 \pm 12</math></b>	<b>72</b>	<b><math>31 \pm 11</math></b>	<b>44</b>	
S_IRON	$70 \pm 19$	$50 \pm 10$	72	$34 \pm 12$	49	
S_NHICE	$68 \pm 19$	$48 \pm 11$	70	$29 \pm 10$	42	
S_REGOLITH	$68 \pm 19$	$48 \pm 11$	70	$26 \pm 9$	39	
S_COM	$70 \pm 18$	$50 \pm 10$	72	$32 \pm 11$	45	
S_SO	$70 \pm 18$	$54 \pm 14$	76	$40 \pm 16$	56	
<b>S_FINAL</b>	<b><math>70 \pm 18</math></b>	<b><math>57 \pm 8</math></b>	<b>81</b>	<b><math>48 \pm 11</math></b>	<b>69</b>	
<b>deep Pacific <math>\delta^{13}\text{C}</math> (‰)</b>						
ODP846 (1.8 Myr)	$0.49 \pm 0.12$	$0.42 \pm 0.17$	86	$0.41 \pm 0.15$	83	
ODP677 (1.3 Myr)	$0.69 \pm 0.09$	$0.57 \pm 0.18$	82	$0.39 \pm 0.30^c$	56	
ODP677/846	$0.55 \pm 0.03$	$0.44 \pm 0.15$	80	$0.40 \pm 0.16$	72	
S_EPICA	$0.54 \pm 0.07$	– <sup>b</sup>	–	–	–	
S_EPICA+	$0.48 \pm 0.07$	– <sup>b</sup>	–	–	–	
<b>S_LR04</b>	<b><math>0.43 \pm 0.12</math></b>	<b><math>0.31 \pm 0.08</math></b>	<b>72</b>	<b><math>0.17 \pm 0.05</math></b>	<b>39</b>	
S_IRON	$0.43 \pm 0.12$	$0.32 \pm 0.07$	73	$0.19 \pm 0.06$	44	
S_NHICE	$0.43 \pm 0.13$	$0.30 \pm 0.08$	71	$0.17 \pm 0.05$	40	
S_REGOLITH	$0.43 \pm 0.13$	$0.30 \pm 0.08$	71	$0.17 \pm 0.05$	39	
S_COM	$0.44 \pm 0.12$	$0.32 \pm 0.07$	73	$0.19 \pm 0.06$	44	
S_SO	$0.43 \pm 0.12$	$0.33 \pm 0.08$	78	$0.22 \pm 0.07$	51	
<b>S_FINAL</b>	<b><math>0.44 \pm 0.12</math></b>	<b><math>0.37 \pm 0.05</math></b>	<b>84</b>	<b><math>0.29 \pm 0.07</math></b>	<b>66</b>	

<sup>a</sup> Variability of ice core  $p\text{CO}_2$  in the MPT contains only three G/IG transitions.

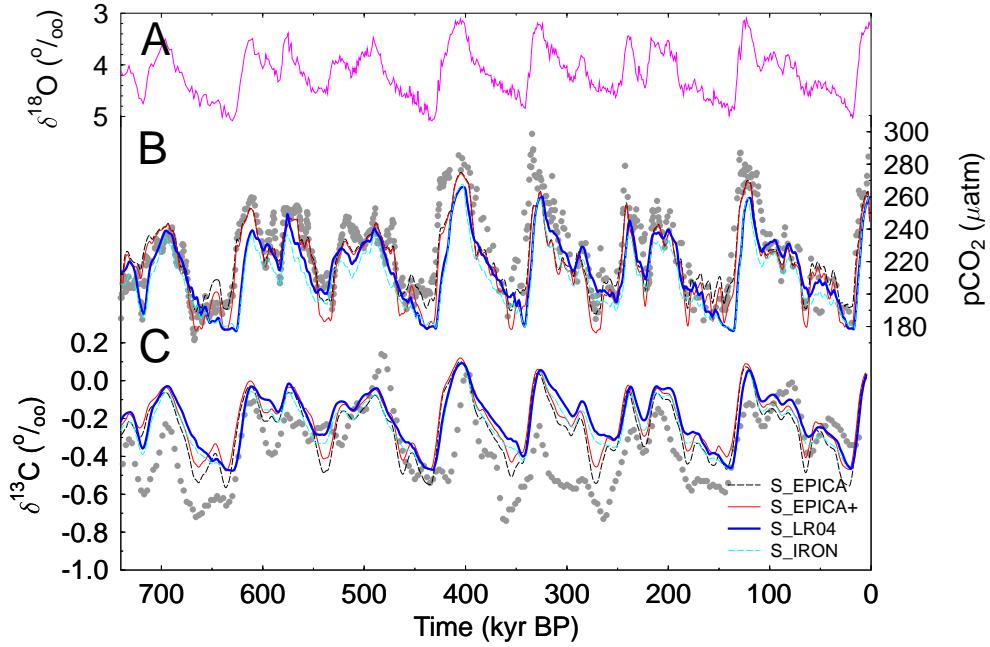
<sup>b</sup> Results of S\_EPICA, S\_EPICA+ are restricted to 740 kyr and are therefore omitted here.

<sup>c</sup> Variability of ODP677 in the 40 k world contains only two G/IG transitions.

tation of  $\delta^{13}\text{C}$  in the Atlantic Ocean (scenario S\_EPICA+). The results of both scenarios in terms of atmospheric  $p\text{CO}_2$  and deep Pacific  $\delta^{13}\text{C}$  are very similar (Fig. 7). Atmospheric  $p\text{CO}_2$  in S\_EPICA agrees very well with the ice core measurements ( $r^2 \approx 0.75$ ). Simulated  $p\text{CO}_2$  in S\_EPICA+ is about  $10\text{ }\mu\text{atm}$  lower during glacial maxima than in S\_EPICA. The simulated deep Pacific  $\delta^{13}\text{C}$  have G/IG amplitudes in the  $100\text{k}$  world ( $0.54\text{‰}$  and  $0.48\text{‰}$  for S\_EPICA and S\_EPICA+, respectively) which agree within their standard deviations with the data-based reconstruction (Table 4).

The approach solely based on the benthic  $\delta^{18}\text{O}$  stack (S\_LR04) leads to atmospheric  $p\text{CO}_2$  which is remarkable

similar to S\_EPICA+. Results of S\_LR04 underestimate  $p\text{CO}_2$  during interglacial periods by  $10\text{ }\mu\text{atm}$  with respect to S\_EPICA+. The original simulations were already failing to reproduce the ice core measurements during these times by about  $20\text{ }\mu\text{atm}$ . This offset is probably based on synchronisation deficits of the individual forcings and neglecting of details on coral reef growth during sea level high stands (Köhler and Fischer, 2006). These biases are less pronounced for the interglacials prior to  $400\text{ kyr BP}$ , for which in the ice cores only moderate atmospheric  $p\text{CO}_2$  values of  $250$  to  $260\text{ }\mu\text{atm}$  are found. During certain short ( $<10\text{ kyr}$ ) time windows the offset between  $p\text{CO}_2$  in S\_LR04 and both the other scenarios



**Fig. 7.** Ground-truthing the simplified forcing approach based on LR04 (comparing it with other scenarios). **(A)** Stacked benthic  $\delta^{18}\text{O}$  (LR04) for comparison. **(B)** Simulated and measured (grey)  $p\text{CO}_2$ . **(C)** Simulated and measured (grey)  $\delta^{13}\text{C}$  in the deep Pacific Ocean. Scenarios are described in Table 2. Data sets as described in Fig. 1.

and the data sets is larger (e.g. 60, 170, 270 kyr BP). These periods were identified to be dominated by an enhanced marine export production (Köhler and Fischer, 2006). As mentioned earlier the correlation between the iron flux to Antarctica (which drives enhanced export production in the Southern Ocean via iron fertilisation) and LR04 is with  $r^2=53\%$  rather poor. Furthermore, the iron flux measured in EPICA Dome C varies over two orders of magnitude. Especially the occurrence and amplitude of peak maxima, which are most important for the marine export production, differs rather strongly between the original ice core data set and its LR04-based surrogate (Fig. 3I). It is therefore not surprising to find that consequences of this process are not depicted very accurately within S\_LR04. Results agree slightly better for the scenario S\_IRON, which uses an alternative regression function between the iron flux and LR04 focused on changes during peaks in the iron flux.

For deep Pacific  $\delta^{13}\text{C}$  the offset between the scenario S\_EPICA+ and S\_LR04 is smaller than for atmospheric  $p\text{CO}_2$  ( $\delta^{13}\text{C}$ :  $r^2=0.75$ ;  $p\text{CO}_2$ :  $r^2=0.62$ ). Especially, there is no systematic bias in  $\delta^{13}\text{C}$  during the last five interglacials as seen in  $p\text{CO}_2$ , however a point-to-point comparison of simulation and reconstruction is due to the missing 500 kyr periodicity in the simulations difficult (see next section for details on this). The disagreements caused by the forcing of the marine export production during short time windows mentioned above is also clearly seen here. The G/IG amplitude in  $\delta^{13}\text{C}$  in the 100 k world reaches with 0.43% for both

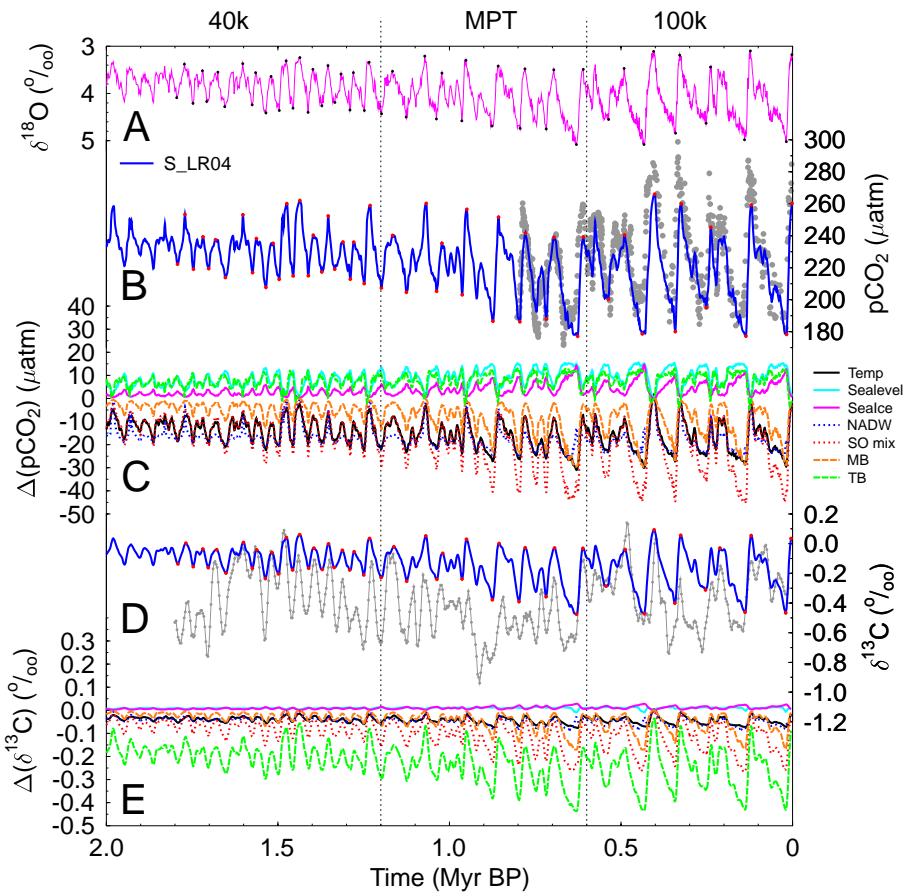
S\_LR04 and S\_IRON about 80% and 90% of what is seen in the ODP cores and the simulation forced with the original paleo records (S\_EPICA+), respectively.

To summarise, our simulation approach based on a simplified forcing of the model with LR04 leads to carbon cycle dynamics which are very similar to the results achieved with the model if forced with the original data sets. About 10% of the G/IG amplitudes in both atmospheric  $p\text{CO}_2$  and deep ocean  $\delta^{13}\text{C}$  are lost through this simplification. Fast features operating on time scales below 10 kyr are not believed to be represented accurately with the LR04-based approach.

### 3.3 The Null Hypothesis for the MPT

We take the comparison presented in the previous subsection as evidence that the general model behaviour based on the simplified forcing approach is in the 100 k world comparable with observations. Therefore, we first test the Null Hypothesis (scenario S\_LR04: climate is similarly related to LR04 before and after the MPT) to interpret the MPT. This implies that no additional processes need to be considered for the interpretation of the carbon cycle during its transition from the 40 k to the 100 k world.

The G/IG amplitudes in atmospheric  $p\text{CO}_2$  are with on average  $31 \pm 11 \mu\text{atm}$  (40 k) and  $51 \pm 12 \mu\text{atm}$  (MPT) much smaller during earlier times than in the 100 k period ( $70 \pm 18 \mu\text{atm}$ ). In the 40 k world  $p\text{CO}_2$  varies only between  $\sim 220$  and  $\sim 260 \mu\text{atm}$ . These amplitudes are rather small,



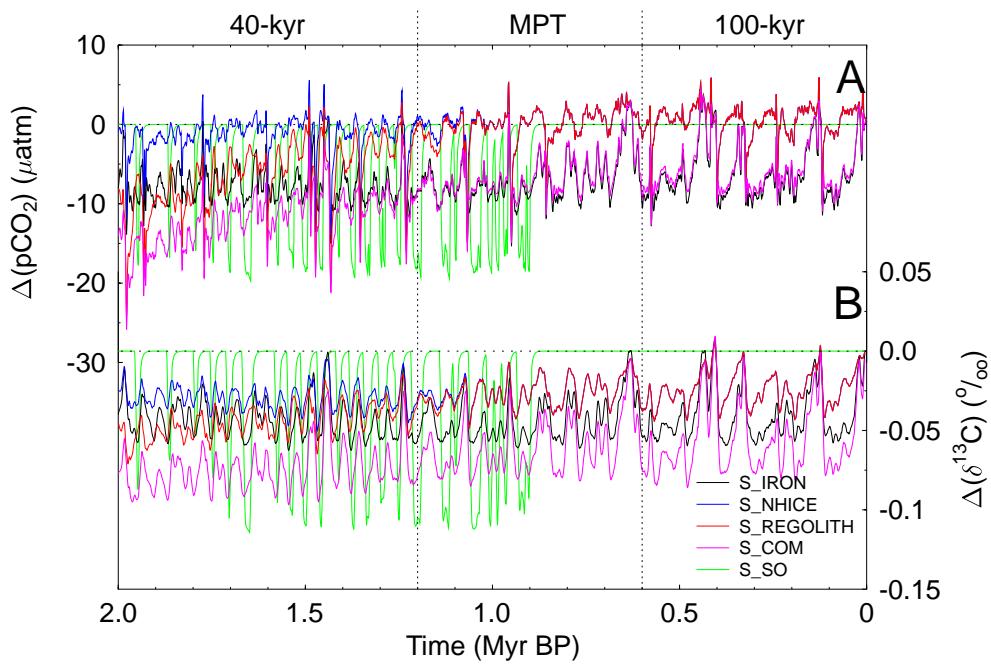
**Fig. 8.** Simulation results of the Null Hypothesis (scenarios S\_LR04). **(A)** Stacked benthic  $\delta^{18}\text{O}$  (LR04) for comparison. **(B)** Simulated and measured (grey)  $p\text{CO}_2$ . **(C), (E)** Contribution of individual processes to changes in  $p\text{CO}_2$  (C) and deep Pacific  $\delta^{13}\text{C}$  (E). The observed processes include changes in ocean temperature (Temp), sea level (Sealevel), gas exchange through sea ice cover (SeaIce), North Atlantic deep water formation (NADW), vertical mixing in the Southern Ocean (SO mix), marine biology due to iron fertilisation of the Southern Ocean (MB) and terrestrial carbon storage (TB). Data sets as described in Fig. 1. Red dots in (B, D) denote local minima/maxima within individual MIS, which were used to calculate G/IG amplitudes.

but we have to consider the known reduction of the G/IG amplitudes of 10% caused by our simplified LR04-based forcing. The relative size of the G/IG amplitudes in  $p\text{CO}_2$  in the 40 k world is 44% of that in the 100 k world. This is smaller than the reduction in the G/IG amplitudes of the climate signal ( $f_{40\text{k}}=51\%$ ) recorded in LR04 (Table 4).

Similarly, the amplitudes in deep Pacific  $\delta^{13}\text{C}$  during the MPT are in scenario S\_LR04 further reduced than in the  $\delta^{13}\text{C}$  data set (Table 4). In the 40 k world they are reduced to only 39% of their 100 k world values, which is about half of the relative size given in the data set (72%), and also smaller than the reduction in the LR04 climate signal (51%). The increase in G/IG amplitude over time in the reconstructions from 0.40‰ (40 k) to 0.55‰ (100 k) is not negligible, indicating already some changes in the carbon cycle. However, this change is reduced by a factor two if only data from ODP846 are considered and are furthermore in the range of

the standard deviation (Table 4). A conservative interpretation of the data might therefore argue for more or less stable G/IG amplitude in deep Pacific  $\delta^{13}\text{C}$  over time. Results of the Null Hypothesis, however, show a rise in G/IG amplitudes by 150% (from 0.17‰ (40 k) to 0.43‰ (100 k)). Based on this disagreement in deep ocean  $\delta^{13}\text{C}$  (see also time series in Fig. 8D) we have to reject our Null Hypothesis (climate is similarly related to LR04 before and after the MPT) to explain the observed variations in the carbon cycle over the MPT. Furthermore, this implies that even the conservative data interpretation (stable G/IG amplitudes over time) asks in the contexts of changing climate as depicted by LR04 for an additional process connected with the MPT, which prevents the carbon cycle from changing.

A spectral analysis (not shown) of the simulated deep Pacific  $\delta^{13}\text{C}$  finds orbital frequencies of about 20, 40, and 100 kyr in the simulation results, similar as in the paleo



**Fig. 9.** Difference between S\_LR04 and alternative scenarios for (A) atmospheric  $p\text{CO}_2$  and (B) deep Pacific  $\delta^{13}\text{C}$ . Scenarios are described in Table 2.

record, but does not find any power in the low frequency component of  $\sim 1/400\text{--}1/500 \text{ kyr}^{-1}$ . This holds for S\_LR04 and all other scenarios discussed below. This discrepancy in the power spectra between model results and reconstruction can be explained if one follows a recent hypothesis on the explanation of the observed  $\sim 500$  kyr cycle in benthic  $\delta^{13}\text{C}$ . According to Wang (2007) it is based on the variability of the monsoon and its long-term impacts on continental weathering and riverine input of bicarbonate into the world ocean and thus the carbon cycle via the hydrological cycle. Because the latter is not included in BICYCLE data and model are expected to disagree in this frequency domain.

The contribution of individual processes to both the variations in atmospheric  $p\text{CO}_2$  and deep Pacific  $\delta^{13}\text{C}$  are identified through a factorial analysis. For this analysis the differences in both variables between the control run (S\_LR04) and simulations in which the one process in question is passive are calculated (Fig. 8C, E). The considered processes here are changes in ocean temperature, sea level, gas exchange rate via sea ice cover, the strength of the Atlantic meridional overturning represented by the NADW formation, vertical mixing and iron fertilisation of the marine biology in the Southern Ocean, and terrestrial carbon storage.  $\text{CaCO}_3$  compensation is active in the whole analysis and the contributions of the individual processes therefore include the partial effect of the sediment/ocean interaction. Processes are thus called to be “equilibrated with the sediments”. This factorial analysis is a first order estimate of individual contributions which neglects nonlinear interacting effects. Some interesting de-

tails can be learnt from this analysis: the variability in deep Pacific  $\delta^{13}\text{C}$  is dominated by changes in terrestrial carbon storage, Southern Ocean vertical mixing and to a certain extent marine productivity. All these three processes contribute to G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  which are larger in the Early than in the Late Pleistocene. Furthermore, the contribution of Southern Ocean processes (vertical mixing and marine export production) to G/IG amplitudes in both variables ( $p\text{CO}_2$ ,  $\delta^{13}\text{C}$ ) was clearly reduced prior to the MPT (Fig. 8C, E). If one seeks a theory which brings simulations of deep Pacific  $\delta^{13}\text{C}$  in better agreement with the reconstructions one might need to revise the temporal changes in one of these processes in the model.

#### 3.4 Alternative scenarios supported by independent evidences

To perform better than our Null Hypothesis alternative scenarios have to produce especially larger G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  before the MPT. Due to the missing  $p\text{CO}_2$  reconstructions in the 40 k world, the performance of simulated atmospheric  $p\text{CO}_2$  is difficult to assess. Differences between the alternative scenarios and the Null Hypothesis (S\_LR04) are summarised in Fig. 9.

*Enhanced marine export production:* If the alternative forcing of aeolian iron flux to Antarctica/Southern Ocean is used (S\_IRON) only small differences to S\_LR04 of up to  $10 \mu\text{atm}$  in  $p\text{CO}_2$  and of  $0.05\text{\textperthousand}$  and Pacific  $\delta^{13}\text{C}$  are found throughout the simulation period. The relative size of the

G/IG amplitudes in the 40 k in comparison to the 100 k world are slightly increased to 49% and 44% for  $p\text{CO}_2$  and  $\delta^{13}\text{C}$ , respectively. The discrepancy between simulated and reconstructed oceanic  $\delta^{13}\text{C}$  is still too large to assess S.IRON as an acceptable scenario.

*Northern Hemisphere glaciation:* The differences of the Null Hypothesis and scenario S.NHICE, which implicitly assumes that changes in sea level are mainly caused by northern hemispheric ice sheets, are with  $<5\ \mu\text{atm}$  for  $p\text{CO}_2$  and  $\sim 0.05\%$  and Pacific  $\delta^{13}\text{C}$  small.  $f$ -ratios are with 42% ( $p\text{CO}_2$ ) and 40% ( $\delta^{13}\text{C}$ ) in the 40 k world very similar to those of the LR04-based scenario.

*Effect of regolith erosion:* Additionally to S.NHICE, the effect of ongoing silicate weathering input through the erosion of the regolith layer beneath the northern hemispheric ice sheets was investigated in S.REGOLITH. The additional and gradually declining input of bicarbonate into the ocean via silicate weathering between 2 and 1 Myr BP (Fig. 3F) leads to a long-term increase in  $p\text{CO}_2$  of about  $10\ \mu\text{atm}$  during the same period of time if compared with S.LR04. This is consistent with other carbon cycle models on chemical weathering (Munhoven, 2002): Higher silicate weathering rates imply a drop in atmospheric  $p\text{CO}_2$ , because not only DIC but also alkalinity in the ocean is changed by the riverine input of bicarbonate. Again, results are very similar to the Null Hypothesis with slightly smaller  $f$ -ratios in the 40 k world.

*Combining all above:* Even if we combine these three alternatives (scenario S.COM), the results are still not improving in a way which leads to G/IG amplitudes in the 40 k world ( $f_{40\text{k}}(\delta^{13}\text{C})=44\%$ ) which are in the range seen in the data sets.

We can therefore summarise, that the improvements achieved through alternative scenarios well supported by other studies are with respect to the simulated G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  in the 40 k world inadequate. None of the alternatives, which are based on either revised forcings due to known weak representation in the Null Hypothesis approach (S.IRON), or on additional evidences how the carbon cycle might have changed during the MPT (S.NHICE, S.REGOLITH), nor a combination of all (S.COM) leads anywhere near the reconstructed variability. We therefore will in the following revise some of our assumptions, in order to suggest another scenario, whose results are in better agreement with the paleo data set.

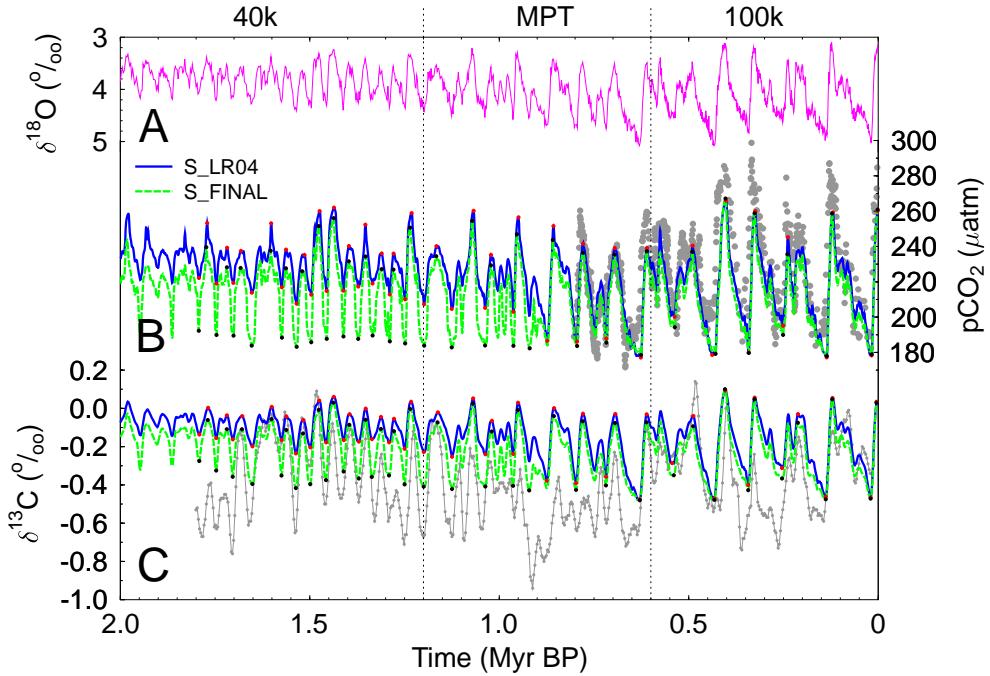
### 3.5 The Southern Ocean Decoupling Hypothesis

One main reason for the use of the LR04-based forcing approach (our Null Hypothesis) is a lack of Antarctic ice core records, which represent Southern Ocean climate, extending 2 Myr back in time. However, there are good reasons to believe, that the extrapolation of LR04-based forcing of SST changes in the Southern Ocean to the 40 k world are reasonable, as reconstructed planktic  $\delta^{18}\text{O}$  and summer SST (Bec-

quey and Gersonde, 2002; Venz and Hodell, 2002) indicate independently and similar to our approach that G/IG amplitudes in SST in the Southern Ocean might have been smaller prior to the MPT (see EPICA Dome C  $\delta\text{D}$  and and substitute in Fig. 3H, which are taken as proxy for Southern Ocean SST). This has in BICYCLE direct consequences for Southern Ocean deep mixing which is a function of SST.

The factorial analysis of the contribution of individual processes to changes in atmospheric  $p\text{CO}_2$  and deep Pacific  $\delta^{13}\text{C}$  (Sect. 3.3, Fig. 8C, E) has shown, that especially the G/IG amplitudes of processes situated in the Southern Ocean differ largely between the 40 k and the 100 k world. Furthermore, these processes have the largest potential, if revised, to bring simulated deep Pacific  $\delta^{13}\text{C}$  in closer agreement to the reconstructions before the MPT. Based on this understanding and the evidences on smaller Southern Ocean SST given above, we suggest, that before the MPT changes in the Southern Ocean vertical mixing rates were decoupled from changes in SST and thus from the global climate change recorded in LR04. This Southern Ocean Decoupling Hypothesis primarily focuses on ocean circulation, but a decoupling of other Southern Ocean processes (e.g. marine export production) from global climate is thinkable. An alternative development of this idea of a decoupling in the Southern Ocean might be that the functional relationship between Southern Ocean SST and the oceanic mixing rates is kept unchanged, but that Southern Ocean SST before the MPT cannot be extrapolated with linear regression functions out of LR04. This might be motivated with the large uncertainty in the G/IG amplitudes in Southern Ocean SST published so far (e.g. Becquey and Gersonde, 2002; Venz and Hodell, 2002). However, this alternative is not developed any further in the following.

The decoupling in the Southern Ocean is proposed because of the impossibility of our carbon cycle model to generate G/IG amplitudes in the 40 k world in deep Pacific  $\delta^{13}\text{C}$  which are comparable with data sets, however first evidences of a different behaviour of the carbon cycle and the climate system emerged already from the analysis of G/IG amplitudes of LR04 and  $\delta^{13}\text{C}$  (Table 4). So far, we have no in-depth idea how the decoupling might operate based on physical principles, but some first estimates how it can be implemented in our carbon cycle model are described in this section and potential physical mechanisms are discussed in Sect. 4. For the time being, we therefore implement in an additional scenario (S.SO) a functional relationship between the Southern Ocean SST surrogate and vertical mixing rates, in which Southern Ocean stratification during glacials is largely enhanced before 900 kyr BP (see Table 3 for details, Fig. 4C). The time for switching this decoupling off was arbitrarily chosen to be exactly the middle of our MPT time window. Other times or a gradually transitions are certainly thinkable. As long as we do not have evidences how this decoupling might operate from first principles we have to acknowledge that this approach here is mainly a first suggestion.



**Fig. 10.** Results for our Null Hypothesis (S\_LR04) and our best guess scenario (S\_FINAL). **(A)** Stacked benthic  $\delta^{18}\text{O}$  (LR04). **(B)** Simulated and measured (grey)  $p\text{CO}_2$ . **(C)** Simulated and measured (grey)  $\delta^{13}\text{C}$  in the deep Pacific Ocean. Data sets as described in Fig. 1. Red and black dots in (B, C) denote local minima/maxima within individual MIS, which were used to calculate G/IG amplitudes.

If the Southern Ocean mixing/stratification is revised as suggested (scenario S\_SO), the carbon cycle variability in the MPT and the 40 k world becomes substantially larger by  $\sim 20 \mu\text{atm}$  and  $0.12\text{‰}$  (Fig. 9). The  $f$ -ratios increase to 78% and 51% for  $\delta^{13}\text{C}$  and 76% and 56% for  $p\text{CO}_2$  for the MPT and the 40 k, respectively. Combining the Southern Ocean Decoupling Hypothesis with all other improvements (scenario S\_FINAL) leads to further increases in the  $f$ -ratios to 84% and 66% ( $\delta^{13}\text{C}$ : MPT or 40 k, respectively) and 81% and 69% ( $p\text{CO}_2$ : MPT or 40 k, respectively). Thus, especially the relative changes in the G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  are now very close to those detected in the ODP records in the MPT (80%) and the 40 k world (72%). In S\_FINAL atmospheric  $p\text{CO}_2$  varies between  $\sim 180$  and  $\sim 260 \mu\text{atm}$  throughout the 40 k world and the MPT (Fig. 10B). This range is similar to the CO<sub>2</sub> observations measured in EPICA Dome C between 450 and 800 kyr BP (Siegenthaler et al., 2005; Lüthi et al., 2008), although the absolute minimum in the ice core data set drops to 170 ppmv in MIS 16. Pacific  $\delta^{13}\text{C}$  is approximately limited to variations between  $0.0\text{‰}$  and  $-0.4\text{‰}$  (Fig. 10C).

Altogether, we can summarise that the Southern Ocean Decoupling Hypothesis together with other improvements of our LR04-based approach leads to simulation results, which are in their relative changes in the G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  comparable with the reconstructions from two ODP records in the equatorial Pacific. So far, simu-

lated  $p\text{CO}_2$  before the MPT varies between 180–260  $\mu\text{atm}$  (Fig. 10). Although these values are a first guess of the variability of atmospheric CO<sub>2</sub> in the 40 k world, they need to be confirmed with either ice core data sets or reconstructions based on oceanic proxies, before they can be taken as reliable. For this comparison of simulation results with paleo reconstructions one has to keep in mind the 10% reduction in G/IG amplitudes of both atmospheric  $p\text{CO}_2$  and deep Pacific  $\delta^{13}\text{C}$  caused by the LR04-based forcing approach.

#### 4 Discussions and conclusions

This study focuses on the understanding of G/IG dynamics in the carbon cycle before, during, and after the MPT. For this aim, simulation results of atmospheric  $p\text{CO}_2$  and deep Pacific  $\delta^{13}\text{C}$  achieved with the carbon cycle box model BICYCLE are compared with reconstructions from sediment and ice cores.

We find that a Null Hypothesis in which climatic changes are mainly represented by the LR04 benthic  $\delta^{18}\text{O}$  stack cannot explain the reduced G/IG amplitudes in Pacific benthic  $\delta^{13}\text{C}$  in the 40 k world. This Null Hypothesis leads to reasonable simulation results in both atmospheric  $p\text{CO}_2$  and deep Pacific  $\delta^{13}\text{C}$  in the 100 k world, in which an independent validation with more complex forced simulations and atmospheric CO<sub>2</sub> data from ice cores is possible. We used this

validation as ground-truthing of the Null Hypothesis, which gave us reasons to believe, that our assumptions might already be sufficient for the interpretation of the carbon cycle during the last 2 Myr. However, even with improvements of our Null Hypothesis with evidences based on other theories the simulated G/IG amplitudes in deep Pacific  $\delta^{13}\text{C}$  are not in agreement with data-based reconstructions. Only if we revise the functional relationship between water column stratification and SST in the Southern Ocean, what we call the Southern Ocean Decoupling Hypothesis, the G/IG amplitudes found in benthic  $\delta^{13}\text{C}$  in the Pacific Ocean can be matched with our simulation results.

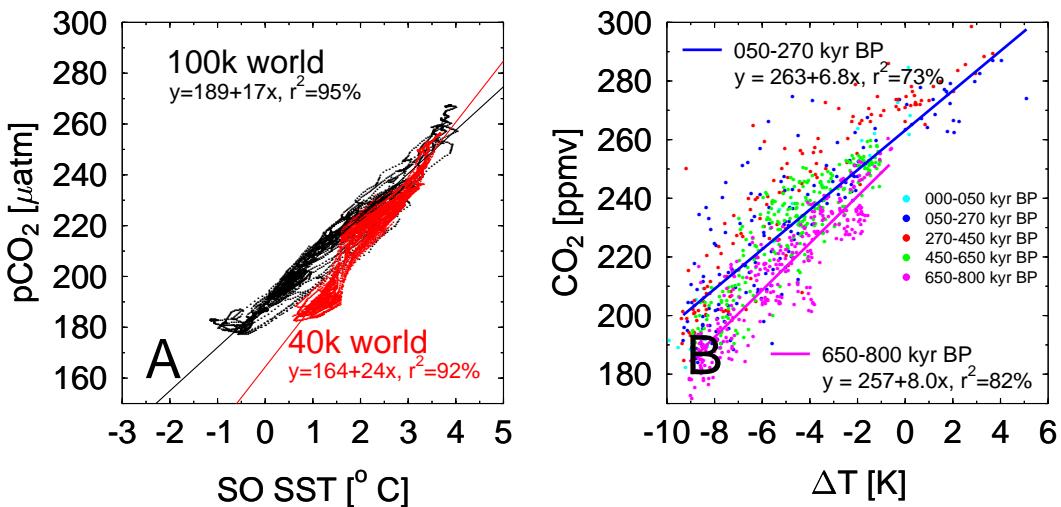
This failure of the Null Hypothesis is remarkable. First evidences for the decoupling of climate and the carbon cycle can already be gained from a comparison of the relative sizes of the G/IG amplitudes before and after the MPT of the paleo records LR04 and benthic  $\delta^{13}\text{C}$  (Table 4). The way how the G/IG amplitudes evolved differently over the MPT in both records (increase by 100% and 40% in LR04 and benthic  $\delta^{13}\text{C}$ , respectively) is a first hint at different dynamics in the climate system and the carbon cycle. Thus, this first evidence of a decoupling is independent of the used carbon cycle model and therefore needs to be addressed to understand the dynamics of the Earth during the MPT regardless of our model-based Southern Ocean Decoupling Hypothesis. It was furthermore surprising to find in the simulation results of our Null Hypothesis (S\_LR04) that in the 40 k world the  $f$ -ratio in  $\delta^{13}\text{C}$  was even smaller than in LR04, thus opposite than in the data sets.

The proposed Southern Ocean Decoupling Hypothesis is additive and not mutually exclusive to those simulations which are based on other theories and which were already combined in one of our scenarios (S\_COM). So far, it seems that where climate evolution over the MPT is of interest, published hypotheses do not consider similar processes as the Southern Ocean decoupling proposed here. Only if the carbon cycle becomes in focus further assumptions on temporal changes have to be considered. From our model-based understanding we believe that the Southern Ocean is a key area also for the interpretation of carbon cycle dynamics over the MPT, as it has already been identified to be a major player for atmospheric CO<sub>2</sub> during G/IG dynamics (e.g. Stephens and Keeling, 2000; Watson and Naveira-Garabato, 2006) and for recent and future uptake of anthropogenic carbon emissions (e.g. Le Quéré et al., 2007; Lovenduski et al., 2007). However, we have to acknowledge that our understanding of the climate and the carbon cycle in the Southern Ocean is incomplete and sometimes contradictory (Toggweiler et al., 2006; Tschumi et al., 2008). In this respect, we argue that those additional information hidden in the proxies of the carbon cycle need more attention and can be used to sharpen theories on climate evolution.

The sensitivity of carbon cycle box models and continuum models was compared in various studies (e.g. Broecker et al., 1999; Archer et al., 2000). It was shown that box models tend to be more sensitive to changes in the high-latitudes than continuum models. Although, this behaviour is especially pronounced in models with three oceanic boxes (Knox and B.McElroy, 1984; Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984), it is also manifested in box models with higher resolution, such as BICYCLE (10 ocean boxes). Note, that BICYCLE's sensitivity to low-latitude changes is in the range of GCMs (for more details see Köhler et al., 2005). Recently, the conclusions of these earlier studies were brought into question. Marinov et al. (2008) analysed in detail how ocean biology affect atmospheric  $p\text{CO}_2$ . They contradict the findings of Archer et al. (2000) and show that one must be extremely careful when comparing different models and when designing comparison indices. According to Marinov et al. (2008) preformed nutrient concentrations are most important for marine biology and for atmospheric  $p\text{CO}_2$ . Small differences in ocean circulation might lead via consequently different preformed nutrient concentrations to large differences in the sensitivity of  $p\text{CO}_2$  to high-latitude forcing.

In this respect it is difficult to judge if or how much of our Southern Ocean Decoupling Hypothesis is dependent on the box model architecture of BICYCLE. The detailed model response is to a certain degree model dependent, as in every simulation experiment, even with more complex models. Various model-intercomparison project have shown that even models of similar complexity behave differently, e.g. the 11 carbon cycle-climate models of the C<sup>4</sup>MIP intercomparison calculated atmospheric CO<sub>2</sub> values at the end of this century ranging between 730 and 1020 ppmv (Friedlingstein et al., 2006). However, more complex models are still not available for the investigations on these long time scales and box models are therefore state-of-the-art for these long integrations.

What independent evidences or alternative theories do we have to support our Southern Ocean Decoupling Hypothesis? Most approaches on the interpretation of the MPT published so far are concentrating on changes in the Northern Hemisphere (e.g. Clark et al., 2007; Bintanja and van de Wal, 2008), although the approach of Bintanja and van de Wal (2008) already considers a 15% contribution of global sea level change from Greenland and Antarctic ice sheets (Bintanja et al., 2002). Recently, Raymo et al. (2006) proposed that changes in climate and ice volume in both the Northern Hemisphere and Southern Hemisphere, each controlled by local summer insolation, need to be considered between 3 and 1 Myr BP for the interpretation of the LR04  $\delta^{18}\text{O}$  stack. During this time window Earth's orbital precession and thus midsummer insolation are out of phase between hemispheres. This implies that 20-kyr changes in ice volume in each hemisphere cancel out in globally integrated proxies such as the oceanic  $\delta^{18}\text{O}$  or sea level leaving the in-phase obliquity (40-kyr) component of insolation to domi-



**Fig. 11.** (A): Correlation between Southern Ocean (SO) SST and atmospheric  $p\text{CO}_2$  in our best guess scenario S\_FINAL. Data and linear regression for the time windows 40 k world (1.8–1.2 Myr BP, red) and 100 k world (0.6–0 Myr BP, black) are shown for comparison. (B): Correlation between Antarctic temperature anomaly  $\Delta T$  and atmospheric  $\text{CO}_2$  as measured in Vostok and EPICA Dome C ice cores. Vostok  $\text{CO}_2$  (Petit et al., 1999) on the orbital tuned age scale (Shackleton, 2000), EDC  $\text{CO}_2$  (Siegenthaler et al., 2005; Lüthi et al., 2008) on the EDC3\_gas\_a age scale (Loulergue et al., 2007; Parrenin et al., 2007b),  $\Delta T$  (Jouzel et al., 2007) on the EDC3 age scale (Parrenin et al., 2007a).  $\text{CO}_2/\Delta T$  data are divided and colour-coded in time windows of two glacial cycles each as in Lüthi et al. (2008).

nate those records. They further hypothesised that before the MPT the extent of the East Antarctic Ice Sheet (EAIS) during interglacials was largely reduced due to higher temperatures. Furthermore, at the MPT marine-based ice sheet margins should have replaced terrestrial ice margins around the perimeter of East Antarctica, resulting in a shift to in-phase behaviour of northern and southern ice volume. The consequence of this “EAIS hypothesis”, that ice shelves might shrink and disappear completely during G/IG cycles in the 40 k world, would certainly have consequences for deep and bottom water formation rates, however, their magnitude is difficult to assess. Existing modelling (Hellmer, 2004) and data-based studies (e.g. Huhn et al., 2008), which focus on the Weddell Sea, highlight that changes in both the ice shelf area and melting at the base of the ice shelf would impact on sea ice and deep ocean properties. In detail, a reduction in both parameters would, according to Hellmer (2004), lead to a decrease in the stability of the water column, enhance deep convection and the formation of dense bottom water. It is difficult to estimate if and how changes in the ice shelf areas around the EAIS would lead to similar consequences. However, based on these studies it can be expected that large changes in Antarctic ice sheet volume and shelf ice extent (as proposed by Raymo et al., 2006) would also impact on ocean circulation. Accordingly, a decoupling of Southern Ocean circulation and SST seems to be a reasonable assumption for the climatic shifts connected with the MPT. The EAIS hypothesis would then imply impacts on Southern Ocean cir-

culation in the 40 k world with the periodicity of the precession (20-kyr). Related to this we cannot detect an increase in spectral power in this frequency band in the 40-kyr world (Fig. 6D–F).

Another theory which focuses on processes in the Southern Hemisphere is the “*insolation canon hypothesis*” of Schulz and Zeebe (2006). It also proposes a connection of the precession cycle and the occurrence of glacial terminations. They analysed the midsummer insolation at  $65^{\circ}$  N and  $65^{\circ}$  S over the last 2 Myr (which both vary mainly with the 20-kyr of the precession) and argued that only when the insolation in both high latitudes increased simultaneously, led by the south, and contained a certain amount of energy a large G/IG transition was triggered. This approach gives hints that insolation changes in both northern and southern high latitudes are important for triggering Terminations I–VII. Further back in time (before the MPT) the insolation regime change: The increase in midsummer insolation was not led by the south or the amount of energy supply corresponding to the insolation increase was smaller than after the MPT and less than a proposed energy threshold necessary for triggering a deglaciation. This is no direct evidence that and how climate in the southern high latitudes differed between Early and Late Pleistocene, but it indicates that the southern high latitudes played a different role during G/IG transitions in the 40 k and in the 100 k world.

Our findings, especially the fact that a decoupling between climate and carbon cycle dynamics is seen in the LR04  $\delta^{18}\text{O}$

and benthic  $\delta^{13}\text{C}$  data without the use of any model, suggests that a revised interpretation of the MPT is necessary, in which changes from the 40 k to the 100 k world in certain processes are required. This conclusion would imply that the interpretation of Huybers (2007) on the MPT might need some refinement. Huybers proposed that there might have been no major abrupt change in the climate system over the MPT. He argued that the glacial variability as seen in a  $\delta^{18}\text{O}$  stack dated independently from orbital forcing can be explained solely by the obliquity cycle given by a 40-kyr periodicity during the last 2 Myr. This hypothesis suggests that the glacial cycles were and are continuously governed by obliquity pacing, but that late Pleistocene glaciations repeatedly skip one or two obliquity cycles, thus resulting in 80 or 120 kyr (on average 100-kyr) periodicity. This would solve both the questions what drove the 100-kyr variability of the late Pleistocene and how the climate system shifted from the 40-kyr towards the 100-kyr variability during the MPT. However, if the decoupling of SST and vertical mixing in the Southern Ocean as proposed in our hypothesis occurred gradually over the last 2 Myr, and not as abrupt as assumed in our scenario S\_SO, then our hypothesis would be in line with the interpretation of Huybers (2007).

We furthermore believe, that hypotheses which concentrate on the Northern Hemisphere glaciation as the dominant process for the interpretation of the MPT alone (Clark et al., 2007; Bintanja and van de Wal, 2008) are not sufficient to explain all the variability seen in various proxy records. They are from our point of view nevertheless valuable and important pieces in the jigsaw to understand the transition from the 40 k to the 100 k world.

We have to point out that due to the simplicity of our LR04-based forcing mechanism and the way we decoupled the Southern Ocean circulation and climate our results are interesting mainly with respect to their G/IG amplitudes. Further details, such as how atmospheric  $p\text{CO}_2$  varied in detail between 180 and 260  $\mu\text{atm}$  are more difficult to assess due to missing  $p\text{CO}_2$  reconstructions in the pre-ice core time window. It has been suggested by various authors (e.g. Clark et al., 2007; Huybers, 2007) that the global cooling trend observed in LR04 might be caused by falling greenhouse gas concentrations. This suggestion is not supported for  $p\text{CO}_2$  from our study, but this might be a model-dependent response, based on the crude implementation of the Southern Ocean decoupling and the coarse representation of weathering fluxes and deep ocean  $\text{CaCO}_3$  chemistry (no process-based model of the sediments included).

Nevertheless, the effect of the Southern Ocean decoupling on atmospheric  $p\text{CO}_2$  would be that the high correlation of Antarctic temperature and atmospheric  $p\text{CO}_2$  found during the last 650 kyr (Siegenthaler et al., 2005) is reduced and that a different relationship between both variables was established in the 40 k world. This is clearly illustrated in the scatter plot of the Southern Ocean SST forcing and the simulated atmospheric  $p\text{CO}_2$  (Fig. 11A), where we find, that the slope

between  $p\text{CO}_2$  and SST is increased in the 40 k world. This means that the same anomaly in Southern Ocean SST would coincide with a larger amplitude in the 40 k world compared to the 100 k world. This trend is already seen if similar relationships are analysed in the ice core records for different time windows (Fig. 11B). If the relation between  $\text{CO}_2$  and Antarctic surface temperature anomaly  $\Delta T$  are analysed, we also find steeper slopes further back in time. Although the differences in the slopes are rather small, they indicate to similar changes in the carbon cycle/climate system over the MPT than our Southern Ocean Decoupling Hypothesis.

In summary, our approach suggests that Southern Ocean climate and ocean circulation might be the variables to look at in the Early Pleistocene and across the MPT. Similar to Wang et al. (2004) we propose that the information contained in the carbon cycle needs to be addressed to get a full picture of Earth's climate variability.

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## Supplemental material to:

# The carbon cycle during the Mid Pleistocene Transition: The Southern Ocean Decoupling Hypothesis

*Climate of the Past, Vol. 4 (2008)*

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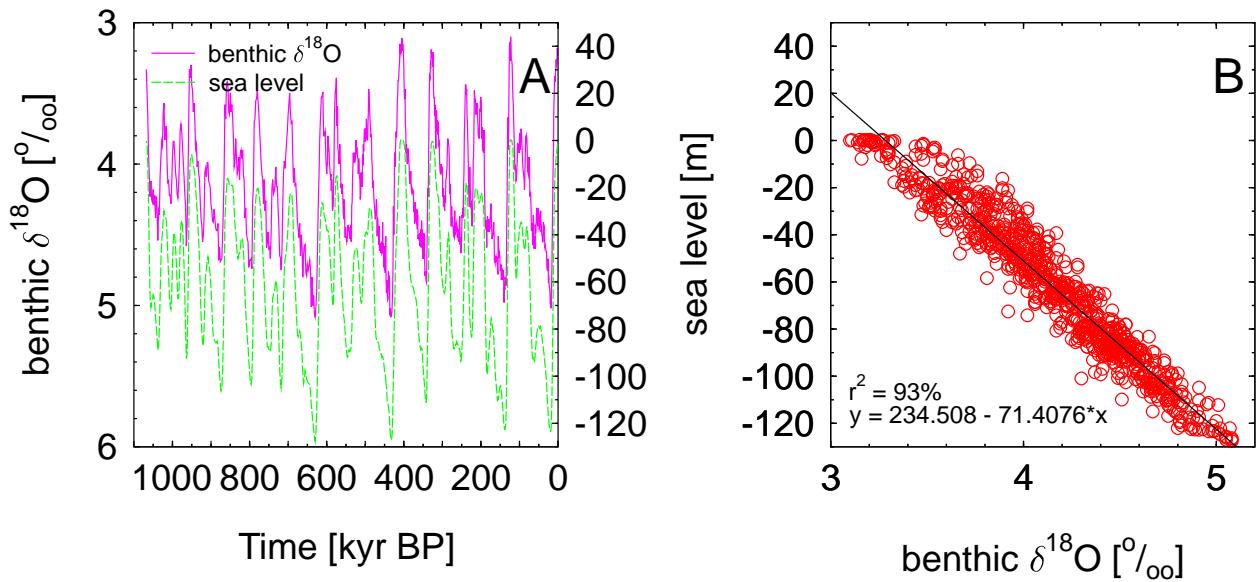
### Appendix A Forcing BICYCLE with LR04

To overcome the problem of synchronisation deficits between various paleo records and because ice core records are so far not extended further back in time than 800 kyr BP, we developed an approach, which is solely based on the LR04 benthic  $\delta^{18}\text{O}$  stack, to force our BICYCLE model over the last 2 Myr. This approach consists of three steps:

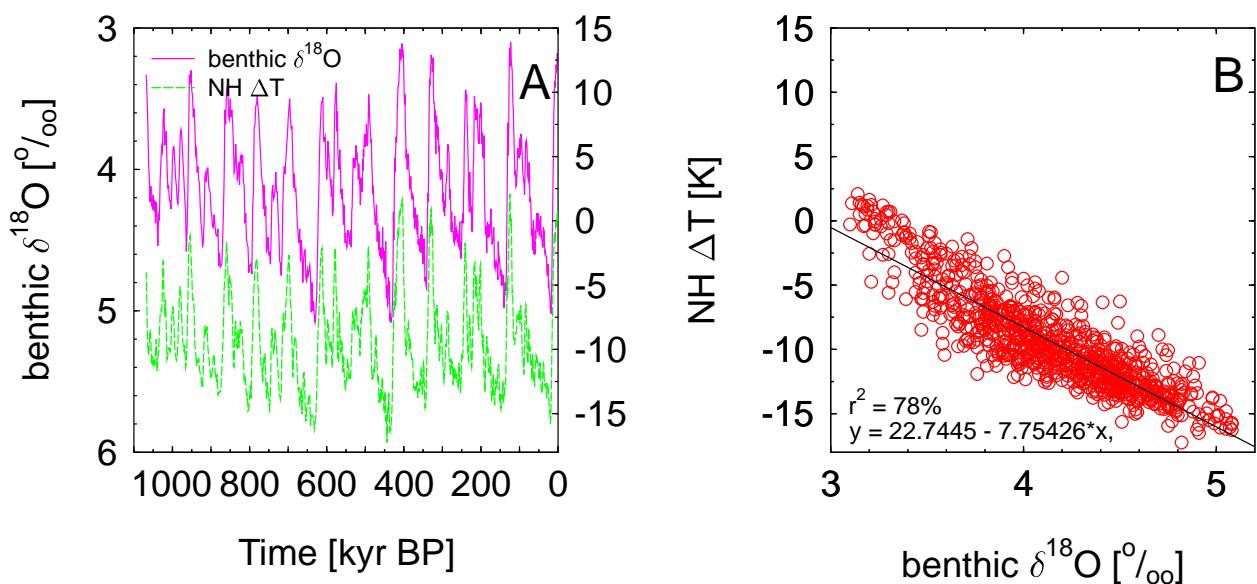
1. Calculate correlations and regressions functions between LR04 and six different climate records which were used in our earlier applications covering the last 740 kyr (Köhler and Fischer, 2006). For this purpose we relied on the available record length and did not restrict the correlation to the previous application period. Details are summarised in Table 1 (main text). Paleo records and regressions used here are plotted in Figs. A1–A6.
2. Decide which and how various climate variables (boundary conditions) of the model depend on the calculated regressions for the approach based on LR04 only (scenarios S\_LR04) and substitute them for the 2 Myr application. Because of the poor  $r^2$  between LR04 and planktic  $\delta^{18}\text{O}$  from ODP677, and because this latter record was available over the whole 2 Myr time period, and its highly synchronisation with LR04, we refrained from substituting this record (which is used as proxy for equatorial SST) at all (Table 3, main text).
3. If necessary (due to inconsistencies or other studies) revise one or more of the substituted climate variables to be used in alternative scenarios (Table 3, main text).

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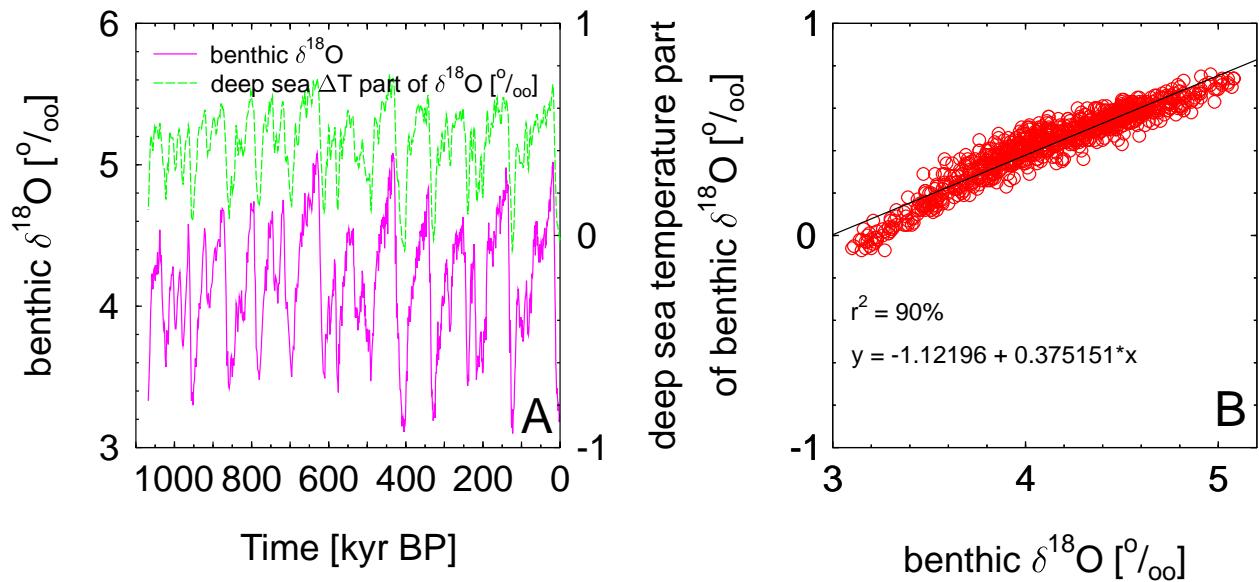
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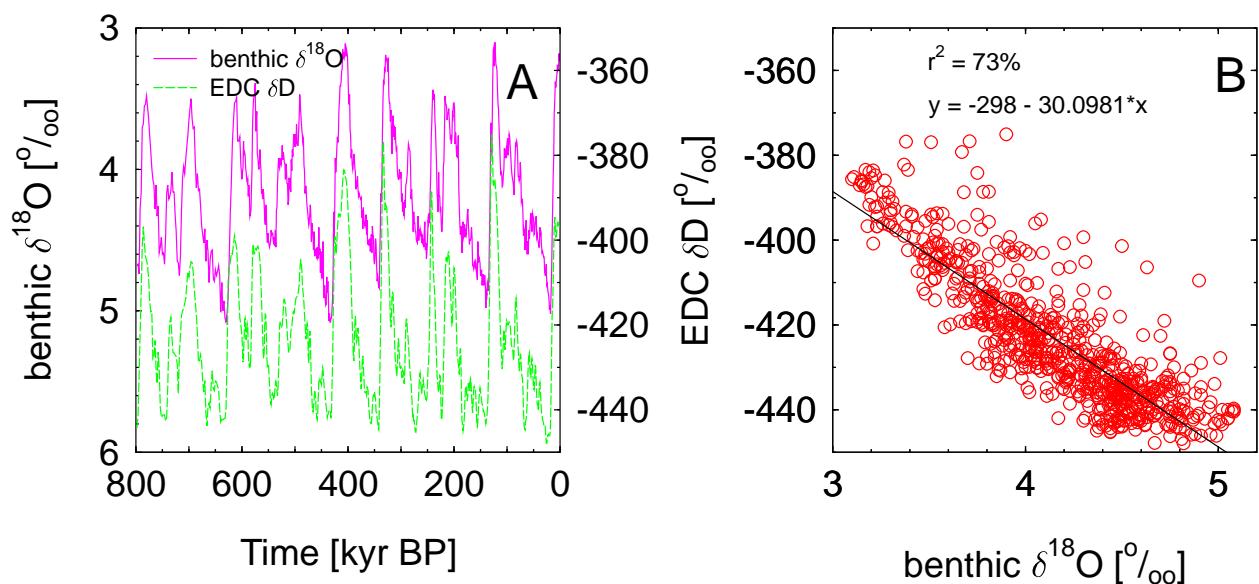
**Fig. A1.** Correlation of stacked benthic  $\delta^{18}\text{O}$  (Lisiecki and Raymo, 2005) with sea level reconstruction of Bintanja et al. (2005). Note the reversed scale of the benthic  $\delta^{18}\text{O}$  stack in A.



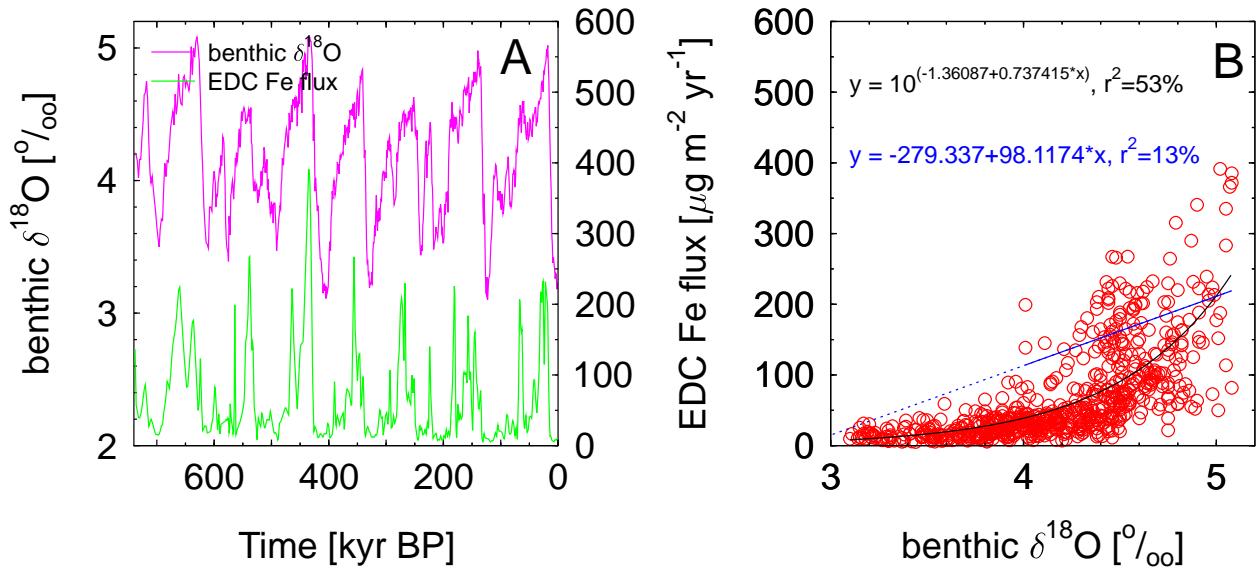
**Fig. A2.** Correlation of stacked benthic  $\delta^{18}\text{O}$  (Lisiecki and Raymo, 2005) with the reconstruction of temperature changes in the northern hemisphere (Bintanja et al., 2005). Note the reversed scale of the benthic  $\delta^{18}\text{O}$  stack in A.



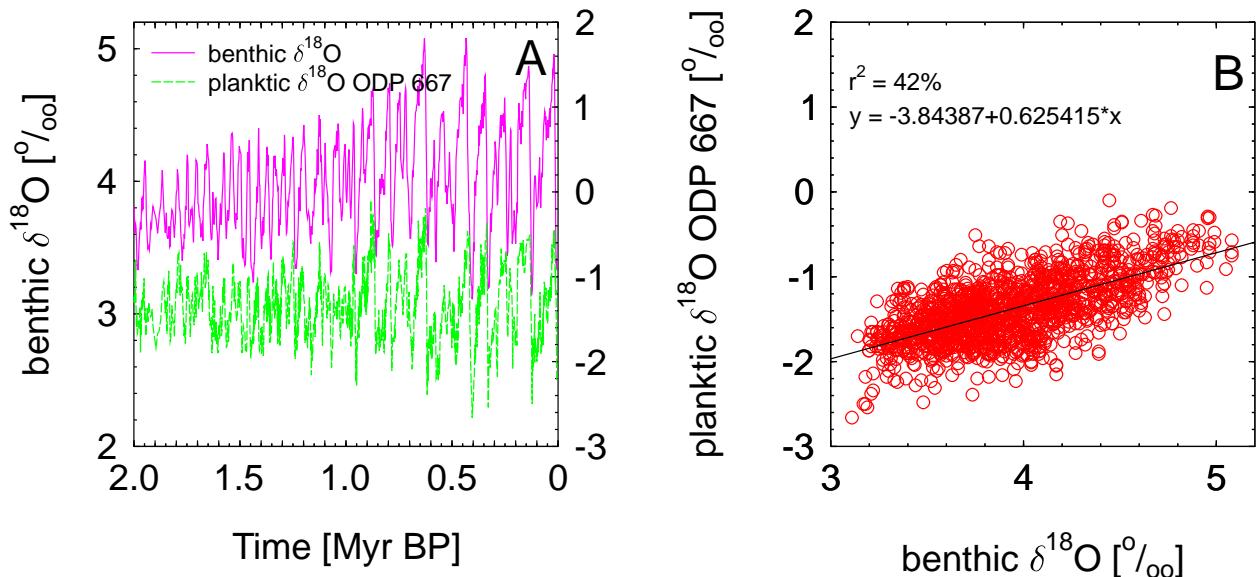
**Fig. A3.** Correlation of stacked benthic  $\delta^{18}\text{O}$  (Lisiecki and Raymo, 2005) with the benthic  $\delta^{18}\text{O}$  part correlated to deep sea temperature changes (Bintanja et al., 2005).



**Fig. A4.** Correlation of stacked benthic  $\delta^{18}\text{O}$  (Lisiecki and Raymo, 2005) with the sea level corrected and smoothed (3 kyr running mean)  $\delta D$  from the EPICA Dome C ice core over the last 800 kyr (EPICA-community-members, 2004; Jouzel et al., 2007) on the EDC3 age scale (Parrenin et al., 2007). Note the reversed scale of the benthic  $\delta^{18}\text{O}$  stack in A.



**Fig. A5.** Correlation of stacked benthic  $\delta^{18}\text{O}$  (Lisiecki and Raymo, 2005) with the Fe flux record from the EPICA Dome C ice core (Wolff et al., 2006) on the EDC3 age scale (Parrenin et al., 2007).



**Fig. A6.** Correlation of stacked benthic  $\delta^{18}\text{O}$  (Lisiecki and Raymo, 2005) with the smoothed (3 kyr running mean) planktic  $\delta^{18}\text{O}$  record ODP667 over the last 2 Myr (Shackleton et al., 1990).