Was the Antarctic glaciation delayed by a high degassing rate during the early Cenozoic?

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ABSTRACT

The Cenozoic is a period of major climatic changes marked by the formation of the Antarctic ice sheet at the Eocene/Oligocene (E/O) boundary. The opening of the southern ocean seaways and the decrease in atmospheric CO2 are two processes generally evoked to explain this E/O cooling. The debate is still ongoing but modeling studies tend to demonstrate that the decrease in atmospheric CO2 is the main driver of the cooling. However, uncertainties persist on what drove the decrease in atmospheric CO2 during the Cenozoic. In this study, we investigate the impact of continental drift, lithology distribution and volcanic degassing rates on the atmospheric carbon dioxide concentration over the Cenozoic within a coupled climate-carbon model (GEOCLIM). In the model, the continental drift results in driving low atmospheric CO2 levels during the Eocene and the Oligocene. The dispersed configuration and the location of a large continental area (North Africa, northern South America) within the Inter Tropical Convergence Zone (ITCZ) promote CO2 consumption by weathering, forcing CO2 to remain low. Icehouse conditions are also promoted by the drifting of India and the weathering of the Deccan basalts in the ITCZ during the Eocene, and by the weathering of the Ethiopian traps during the Oligocene. To prevent the building up of the Antarctic ice sheet at the Eocene, the model needs enhanced solid Earth degassing flux by 50% so that atmospheric CO2 levels stay above the glacial threshold (750 ppm). We find that the decrease in atmospheric CO2 from the Eocene to the Oligocene is probably due to a reduction in the source of volcanic CO2 rather than an increase in silicate weathering. The model results furthermore suggest that during the Miocene period, the northward drifting of both the African plate and India (including the Deccan traps) might have decreased the continental surface exposed to chemical weathering, therefore generating higher CO2 values. Finally, the uplift of the Tibetan plateau from the Miocene to the present-day induces in the model an increase in silicate weathering through the intensification of the South-Eastern Asian monsoon, causing atmospheric CO2 to come back to a pre-industrial value.

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1. Introduction

It is well documented that the Cenozoic is a period of transition from the warm Cretaceous climate toward the cooler present-day (Zachos et al., 2001). The two most salient features of the Cenozoic are the growth of the ice sheet over Antarctica at the Eocene–Oligocene boundary, and the onset of the Northern hemisphere glaciation about 5 million yr ago. For both of these events, a threshold of atmospheric CO2 to initiate glaciation has been estimated to be 750 ppm for Antarctica (DeConto and Pollard, 2003), and about 280 ppm, a much lower level, for the Northern hemisphere (DeConto et al., 2008). Any reconstructions of the long-term CO2 evolution during the Cenozoic must account for these two threshold values and respective timing.

Until now, the most commonly invoked mechanism for the Cenozoic CO2 drawdown has been the reduction of CO2 by weathering of silicate rocks promoted by the Himalayan uplift (Raymo and Ruddiman, 1992). However the weathering of silicate rocks exposed in the Himalayan orogen is alone not large enough to account for the observed drawdown in atmospheric CO2 (France-Lanord and Derry, 1997; Galy and France-Lanord, 1999). The associated burial of organic carbon in the Bengal fan might have helped to contribute in drawing CO2 down (Galy et al., 2007). This organic carbon sink is quite efficient today and throughout the Neogene, (Derry and France-Lanord, 1996). Other orogens could
have been additional important contributors to the Cenozoic CO2 drawdown. Moquet et al. (2011) showed that silicate weathering in the Andes substantially contributes to the CO2 consumption today. Furthermore, weathering may take place in floodplains at the foot of active orogens (Bouchez et al., 2012). Despite years of debate, we have to acknowledge that, up to now, no quantitative model describing the integrated effects of mountain uplift on the carbon cycle evolution has been proposed. The reason for this is rooted in the complexity of the processes at play. In particular, a process-based description of the time evolution of the mechanical erosion is still lacking.

Another mechanism for the Cenozoic cooling is the general decrease in the solid Earth degassing rates, as shown by Gaffin (1987) and Engebretson et al. (1992), who suggested that ocean crust production has halved since the Cretaceous. On the contrary, Larson (1991) and Rowley (2002) suggested that the overall degassing rate showed little variation over the last 100 Ma. According to these reconstructions, the past solid Earth degassing rate is still largely unconstrained, despite its central role in the evolution of global climate during the Cenozoic.

A third mechanism for the atmospheric CO2 evolution is better constrained over the Cenozoic and refers to the continental drift. When moving either in latitude or in longitude, the continents go across various climatic belts. These displacements strongly influence the chemical weathering of rocks. Marshall et al. (1988) were the first to propose the existence of a causal link between the continental configuration, the silicate rock weathering and the atmospheric CO2 level, using a simple zonally averaged model of the land distribution. Bluth and Kump (1991) and Gibbs et al. (1999) showed with a GCM, and a realistic continental drift history, that the end Mesozoic–early Cenozoic periods have been characterized by an efficient silicate weathering promoted by intense rainfall. More recently, the continental drift effect on atmospheric CO2 and global climate has been explored with the GEOCLIM model for the whole Phanerozoic (Donnadieu et al., 2006; Goddéris et al., 2008, 2012; Nardin et al., 2011). Superimposed on the continental drift, several large igneous provinces (LIPs) erupted during the Cenozoic or slightly before (Deccan traps, Ethiopian traps; Courtillot and Renne, 2003). Together with older basaltic provinces, and because they weather much faster than continental shield rocks (Dessert et al., 2003), LIPs constitute potential hot spots for CO2 consumption when carried by continental drift into warm and humid areas (Kent and Muttoni, 2008; Lefebvre et al., 2010).

In this contribution, we use the coupled climate-carbon model GEOCLIM to quantify the impact of the Cenozoic drift of the continents and the LIPs on the time evolution of the atmospheric CO2.

2. Model description

2.1. General framework of the GEOCLIM model

We use the most recent version of the GEOCLIM model described by Donnadieu et al. (2009). The model couples the 3-D climate model FOAM, the dynamic global vegetation model LPJ (Sitch et al., 2003) and the box model for the geological carbon and alkalinity global cycle COMBINE (Goddéris and Joachims, 2004).

The atmospheric component of FOAM is a parallelized version of NCAR's Community Climate Model 2 (CCM2) with the upgraded radiative and hydrologic physics incorporated in CCM3 v. 3.2. The atmosphere runs at R15 spectral resolution (4.5° × 7.5°) with 18 levels. For this study, we use FOAM in mixed-layer mode where the atmosphere is coupled to a 50-m mixed-layer ocean, which parameterizes heat transport through diffusion, in order to save computation time (one GEOCLIM simulation requires up to 12 GCM simulations).

The vegetation model LPJ is a Dynamic Global Vegetation Model of vegetation biogeography and vegetation/soil biogeochemistry. Starting from the climate, soil and atmospheric information as input, it dynamically computes the transient vegetation composition in terms of plant functional groups and their associated carbon and water budgets. The calculated vegetation distribution directly impacts on (1) the albedo of each grid cell, (2) its thermal properties, (3) its roughness, and (4) its potential evaporation.

In this contribution, we aim to calculate the steady-state pCO2 associated with five continental configurations spanning the whole Cenozoic. On a geological timescale, the global carbon and alkalinity cycles are balanced:

\[ F_{vol} = F_{silw} + (F_{ad} - F_{ow}) \]  \hspace{1cm} (1)

where \( F_{vol} \) is the total solid Earth CO2 degassing flux and \( F_{silw} \) is the total CO2 consumed by continental silicate rocks (Walker et al., 1981). \( F_{ad} \) is the burial of organic carbon and \( F_{ow} \) the oxidation of sedimentary reduced carbon exposed at the continental surfaces. The difference, \( F_{ad} - F_{ow} \), fluctuated over the Cenozoic (Goddéris and Franco, 1996). However, because in the present study we focus on estimating the role of continental drift and its impact on silicate weathering on the Cenozoic climate evolution, we deliberately set the \( F_{ad} - F_{ow} \) term to 0. This assumes steady state of the organic carbon sub-cycle. Under this assumption, Eq. (1) simplifies as follows:

\[ F_{vol} = F_{silw} \]  \hspace{1cm} (2)

\( F_{silw} \) depends on climate (runoff and air temperature), amongst others, and climate depends on atmospheric CO2 pressure. Accounting for these dependencies, Eq. (1) is readily solved for the steady-state atmospheric CO2.

The advantage of GEOCLIM when compared to previously published models such as GEOCARB (Berner, 2004) is to resolve the continental climate spatially, so that climatic parameters such as runoff, air temperature and consumption by continental silicate weathering, are calculated at each grid cell. We can rewrite Eq. (1) including the impact of the continental configuration:

\[ F_{vol} = \sum_{j=1}^{ncont} F_{silw}(T_j, \text{runoff}_j) \]  \hspace{1cm} (3)

where \( ncont \) is the number of continental grid elements, \( T_j \) and \( \text{runoff}_j \) the air temperature and runoff calculated by the GCM for the grid element \( j \) respectively.

FOAM is run for each paleogeographical configuration at various CO2 concentrations (from 160 to 2400 ppm) accounting for the secular trend in the solar constant. Each simulation is run for 30 yr (until steady state is reached). Assuming that the modeled climate system behaves quasi-linearly, we can estimate the climatic variables (temperature and runoff) for any pCO2 and continental configuration to calculate the \( F_{silw} \) term on each continental grid cell. Then, if \( F_{vol} \) is known, we can estimate the atmospheric CO2 level necessary for equilibrating \( F_{vol} \) and \( F_{silw} \) (Eq. (2)).

2.2. Model representation of the lithology

The geochemical model differentiates weathering of basaltic and granitic lithologies. For the granitic lithologies, total CO2 consumption through weathering is based on the work by Oliva et al. (2003), who compiled data for about 100 small granitic catchments (Eq. (4)). They showed that the cation flux released by granitic weathering is proportional to the runoff and is a function of temperature, assuming an apparent activation energy of 48.7 kJ/mol. The CO2 consumption is proportional to the cation...
flux, because in freshwater the cationic charge is mostly balanced by HCO$_3^-$ anions. CO$_2$ consumption is then proportional to the runoff and is a function of temperature with the same activation energy as the total cation flux.

$$F_{\text{grow}} = \sum_{j=1}^{npixel} k_{\text{grow}} \cdot \text{runoff}_j \cdot \text{area}_j \cdot \exp \left[ \frac{E_{\text{grow}}}{R} \left( \frac{1}{T} - \frac{1}{T_0} \right) \right]$$ (4)

where $F_{\text{grow}}$ is the total atmospheric CO$_2$ consumption via granitic weathering, $R$ is the perfect gas constant, $E_{\text{grow}}$ is the apparent activation energy of 48.7 kJ/mol, $\text{area}_j$ is the surface of the continental grid cells, $T_j$ and $\text{runoff}_j$ are the mean annual ground air temperature and the annual mean runoff at the grid cell $j$ respectively.

Weathering of basaltic lithologies is calculated according to Dessert et al. (2001):

$$F_{\text{basw}} = \sum_{j=1}^{npixel} k_{\text{basw}} \cdot \text{runoff}_j \cdot \text{area}_j \cdot \exp \left[ \frac{E_{\text{basw}}}{R} \left( \frac{1}{T} - \frac{1}{T_0} \right) \right]$$ (5)

where $F_{\text{basw}}$ is the total atmospheric CO$_2$ consumption flux through basaltic weathering. $E_{\text{basw}}$ is fixed at 42.3 kJ/mol. $k_{\text{basw}}$ and $k_{\text{grow}}$ are calibrated on present day climatic conditions provided by a FOAM simulation in a mixed-layer configuration. By assuming a total CO$_2$ consumption by silicate weathering ($F_{\text{grow}}+F_{\text{basw}}$) equal to 13.6 x 10$^{12}$ mol/yr (Dessert et al., 2003; Gaillardet et al., 1999) with a 30% contribution of basaltic lithologies and by using the present day climatic fields from FOAM, $k_{\text{basw}}$ (6.25 x 10$^3$ with runoff in cm/yr and area in 10$^6$ km$^2$) and $k_{\text{grow}}$ (1.23 x 10$^9$ with runoff in cm/yr and area in 10$^6$ km$^2$) are determined.

2.3. Model configurations

We use five geographic reconstructions of the Cenozoic including Early Paleocene (65 Ma), Early Eocene (52 Ma), the middle of the Oligocene (30 Ma), Middle Miocene (15 Ma) and Present-Day (PD). The Miocene geography comes from Herold et al. (2008). The land–ocean configurations for 65 Ma, 52 Ma and 30 Ma are built from a synthesis of paleomagnetic data, hot spot tracks and geologic constraints (Besse and Courtillot, 2002; Dercourt et al., 1993).

Earth’s orbital parameters are set to present-day values for the five configurations. Solar radiation is assumed to evolve through time according to the stellar evolution models (from 0.65% reduction in the Paleocene to 0.15% in the Miocene, Gough, 1981). We use the estimated CO$_2$ thresholds below which ice is allowed to accumulate on Antarctica and Greenland at 750 ppm (DeConto and Pollard, 2003) and 280 ppm (DeConto et al., 2008; Lunt et al., 2008) respectively. Based on these estimations, and because the FOAM GCM does not include an ice sheet model allowing the onset of an ice sheet, (1) we assume the absence of ice sheets in all geographies when CO$_2$ is above 750 ppm, (2) we prescribe the existence of both ice sheets (North and South) for CO$_2$ levels lower or equal to 280 ppm, and (3) we prescribe the Antarctic ice sheet for CO$_2$ levels between 280 and 750 ppm. The ice sheet relief and albedo values for Antarctica and Greenland are taken from their modern states.

2.4. Setup of the atmospheric CO$_2$ simulations

Over the Cenozoic, mechanical erosion appears to increase worldwide in response to the erosion of the major Cenozoic orogens (Kump and Arthur, 1997; Wold and Hay, 1990). In order to account for a fluctuating role of physical erosion in our simulations, potentially promoting silicate weathering (West et al., 2005), we simply multiply our granitic silicate weathering rate by a $f_R$ factor which varies from 0.75 at 65 Ma to 1 at 15 Ma (method and values are from Berner (2004)). We then run three sets of simulations. In the first one (UNI), we calculate the steady state CO$_2$ for each time period assuming a constant degassing rate ($6.8 \times 10^{12}$ mol/yr). A uniform distribution of the basaltic rocks at each terrestrial grid cell is prescribed and calibrated in order for the basalt weathering to reach a 30% contribution of the total silicate weathering flux taken at present day (Dessert et al., 2003). These simulations aim to show the effect of the continental drift. In the second set of simulations (REG simulation), the consumption of CO$_2$ by continental silicate weathering is calculated based on the spatial distribution of the basaltic provinces. Fig. 1 shows prescribed distribution of basaltic provinces plotted in the atmospheric resolution of the model FOAM. The size of each basaltic province is listed in Table 1. In the REG simulation, solid Earth degassing rate is fixed to present day value. In the third set of simulations (VOL), igneous provinces are explicitly accounted for as in the REG simulation, but we now consider a time-varying degassing rate throughout the Cenozoic. This is deduced from the reconstruction of the subduction rate provided by Engelbreton et al. (1992) (see discussion below). These simulations are summarized in Table 2.

3. Results

3.1. Role of the continental drift

The most striking result comes from the UNI simulations where atmospheric CO$_2$ stays relatively constant oscillating around 340 ppmv over the whole Cenozoic (Fig. 2; Table 3). This is not

<table>
<thead>
<tr>
<th>Basaltic provinces</th>
<th>Age (Ma)</th>
<th>Area ($10^6$ km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kamchatka (Kam.)</td>
<td>5-12</td>
<td>0.17</td>
</tr>
<tr>
<td>Patagonia (Pat.)</td>
<td>10</td>
<td>0.21</td>
</tr>
<tr>
<td>Columbia River (Col.)</td>
<td>16</td>
<td>0.2</td>
</tr>
<tr>
<td>Ethiopia (Eth.)</td>
<td>30</td>
<td>0.8</td>
</tr>
<tr>
<td>Island/Greenland (Isl.)</td>
<td>60</td>
<td>0.1</td>
</tr>
<tr>
<td>Deccan (Dec.)</td>
<td>65</td>
<td>0.5</td>
</tr>
<tr>
<td>Central America (Cent. Am.)</td>
<td>90</td>
<td>0.31</td>
</tr>
<tr>
<td>Indonesia/SE Asia (Indo.)</td>
<td>90-250</td>
<td>0.54</td>
</tr>
<tr>
<td>Parana (Par.)</td>
<td>133</td>
<td>0.57</td>
</tr>
<tr>
<td>Karoo (Kar.)</td>
<td>184</td>
<td>0.22</td>
</tr>
<tr>
<td>Siberia (Sib.)</td>
<td>250</td>
<td>1.5</td>
</tr>
<tr>
<td>Emeishan (Emei.)</td>
<td>259</td>
<td>0.5</td>
</tr>
<tr>
<td>Timan/NW Russia (Tim.)</td>
<td>550</td>
<td>0.18</td>
</tr>
<tr>
<td>Tasmania (Tas.)</td>
<td>580</td>
<td>0.33</td>
</tr>
</tbody>
</table>

Total 6.13

Fig. 1. Present-day localization of the main basaltic provinces and their age of formation. Comparing the area of an igneous province with the area of the model grid cell, we define the fraction of the cell covered by the igneous province (the rest being covered by shield rocks).

Table 1

Ages and areas of the main basaltic provinces considered in this study (Chen et al., 2008; Corgne et al., 2001; Courtillot and Renne, 2003; Dessert et al., 2003; Honthaus et al., 1995; Kuznetsov et al., 2007; Meffre et al., 2004; Wignall, 2001).
Table 2
Description of the parameters used in the simulations.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Basalt repartition</th>
<th>Mechanical erosion (65–15 Ma)</th>
<th>Degassing rate</th>
<th>Aims</th>
</tr>
</thead>
<tbody>
<tr>
<td>UNI</td>
<td>Uniform</td>
<td>0.75–1</td>
<td>Constant (6.8 × 10^12 mol/yr)</td>
<td>Continental drift</td>
</tr>
<tr>
<td>REG</td>
<td>Explicit with the modern geographical distribution (Fig. 1)</td>
<td>0.75–1</td>
<td>Constant (6.8 × 10^12 mol/yr)</td>
<td>Continental drift + basaltic</td>
</tr>
<tr>
<td>VOL</td>
<td>Explicit with the modern geographical distribution (Fig. 1)</td>
<td>0.75–1</td>
<td>Deduced from Engebretson et al. (1992) (Fig. 51)</td>
<td>Degassing rate</td>
</tr>
</tbody>
</table>

because the carbon cycle is insensitive to the change in continental configuration, but rather the result of efficient weathering during the Cenozoic damping CO₂ down to low levels. Indeed, using the same model, Donnadieu et al. (2009) found higher CO₂ levels for pre-Cenozoic continental configurations. In addition, subtle changes in the paleogeography of our model such as the northward drift of Pangea between the Carnian and the Rhetaen time periods (i.e. 220 and 200 Ma) can trigger an atmospheric CO₂ fluctuation of 2000 ppm (Goddéris et al., 2008). To illustrate the effect of paleogeography in our model, we compare global weathering rates from climatic simulations with identical atmospheric CO₂ level (280 ppmv, i.e. the pre-industrial value) (Fig. 3). In these series of simulations, the basaltic weathering rates are substantially larger in the 65, 52 and 30 Ma configurations and lower in the 15 Ma configuration than the modern value (Fig. 3b). For the 65, 52 and 30 Ma paleogeographies, the trend in granitic weathering rates is slightly different because of the erosion corrective factor that implicitly draws down the weathering rates of the granitic surfaces by 25% (Fig. 3a). As a result, the global weathering rate, the sum contribution of basaltic and granitic lithologies, stays below (but close to) the present-day level (Fig. 3c). With lower global weathering rates, atmospheric CO₂ levels stay above the pre-industrial value in the model (Fig. 2).

In this set of experiments, we did not aim to quantify the effect of the physical erosion of active orogens on the global carbon cycle. However our results suggest that the weathering feedback remains quite efficient during the early Cenozoic even in the absence of intense physical erosion.

Modeled CO₂ levels for the 65 Ma and 15 Ma time periods are well within the range of estimates, while modeled CO₂ levels of the Oligocene and Eocene are at the lower end estimates (Fig. 2). However, all CO₂ levels from the UNI simulations are lower than the threshold for initiating an extensive glaciation in Antarctica. In these simulations, the transition to an icehouse world would be triggered from the beginning of the Cenozoic. Because the main mechanism in the UNI simulation is the continental drift forcing the global climate to be cold well before the effective start of the glaciation, we conclude that another mechanism is required to explain the evolution of the long-term CO₂ trend (and climatic trend) throughout the Cenozoic.

3.2. Role of the weathering of the LIPs

The REG simulation (where basaltic provinces are spatially distributed) shows larger fluctuations in the atmospheric CO₂ levels than the UNI simulation. In REG, atmospheric CO₂ concentration stays relatively high during the Paleocene (680 ppm) and the Miocene (500 ppm), and drops to about 340 ppm at the Eocene and Oligocene (Fig. 2; Table 3). To separate the different causes in the REG CO₂ trend, we calculate the silicate weathering flux for each reconstruction set with 280 ppm of CO₂ (Figs. 3 and 4). The model results show that the surface of basaltic provinces experiencing a warmer and wetter climate is smaller during the Paleocene than during the rest of the Cenozoic. This is because the Ethiopian traps have not been emplaced at that time, and the 258 Ma-old Emeishan and 65 Ma-old Deccan traps are still located in subtropical regions where runoff is low, inducing smaller global silicate weathering rates (4.87 × 10^12 mol of CO₂/yr for the Paleocene vs. 6.8 × 10^12 mol of CO₂/yr for the modern, see Figs. 3 and 4). The lower basaltic weathering rate at the Paleocene is compensated in the model by higher atmospheric CO₂ concentration in order to keep global silicate weathering and total solid Earth degassing in balance (Eq. (1)). It is only during the Eocene that the Deccan traps finally leave the subtropical latitudes to enter the ITCZ increasing basaltic and silicate weathering (Fig. 4). An additional contribution at the Oligocene to the global silicate weathering comes from the weathering of the Ethiopian traps below the ITCZ (Figs. 3 and 4). During the Miocene, the northward drift of Africa and India reduced the basaltic weathering over the Ethiopian traps and the Deccan traps respectively, resulting in higher CO₂ concentration while reaching steady state in the model (Figs. 3 and 4). Finally, from the Mid-Miocene to the present day, the silicate weathering flux increases in response to the strengthening of precipitation and runoff in southeastern China and Indonesia due to the Tibetan plateau uplift and subsequent intensification of the South East Asian monsoon (Fig. 4). Observations show a more intense monsoon in Asia during the last 8–9 Ma (Zhisheng et al., 2011). The displacement of the Emeishan traps and the Indonesian basaltic archipelago in the vicinity of the ITCZ emphasizes this pattern of increased weathering over the last 15 Ma. The modeled evolution in the weathering of the main basaltic provinces, calculated for each time slice of the REG simulation, confirms this scenario (Fig. 5). The model shows that weathering specifically of the Deccan traps reaches a maximum at 52 Ma, followed by a decrease during the Oligocene to reach relatively low rates from the Miocene to today. The Ethiopian traps are rapidly weathered during the Oligocene, directly after their onset, and consume less CO₂ during the Miocene and the present-day. Finally, our model simulations show that CO₂ concentration levels kept decreasing in the last 15 Ma because of a strengthening of the East Asian monsoon that resulted in stronger weathering of Emeishan and Indonesia basaltic provinces.

The explicit calculation of the effect of the basaltic provinces leads to a general increase in CO₂ from the UNI to the REG simulations. Nevertheless, atmospheric CO₂ remains too low, i.e. below the Antarctic glaciation threshold, during the Paleocene (680 ppm) and the Eocene (360 ppm).

3.3. Role of the degassing rate

The evolution of the global degassing rate throughout the Cenozoic, and more generally throughout the Earth history, remains poorly constrained. There are (1) no direct measurements of the Cenozoic fluctuations of the solid Earth degassing and (2) no clear consensus between the various reconstructions of the degassing flux through the geological past. For instance, recent re-evaluations suggest that because the ridge production is estimated to have stayed roughly constant for the last 180 Ma, the degassing rate should have had little fluctuations since the middle Jurassic (Rowley, 2002). However, as already noticed by Kerrick and Caldeira (1999), processes other than ridge production can influence
the Earth degassing rate. During the Paleocene and the Eocene, Kerrick and Caldeira (1999) suggested that the Earth degassing rate may have increased in response to regional metamorphism of the Tethysian orogeny, following the subduction of carbonates accumulated on the seafloor of the European and Asian margins. Using this information along with Engebretson et al.'s (1992) work on subduction history, we estimate that the global degassing rate was higher between 130 and 50 Myr than today (Fig. A1). We investigate here the potential impact of this degassing history on modeled atmospheric CO2. Increasing the degassing rate by 50% results in the potential impact of this degassing rate history on modeled atmospheric CO2. Increasing the degassing rate by 50% results in the long-term CO2 calculated in the VOL simulation is in good agreement with the evolution of the climate during the Cenozoic. The intensity of weathering during the Paleocene and Eocene requires an enhanced CO2 source (Engebretson et al., 1992) to allow a greenhouse climate that prevents the Antarctic glaciation, which might have happened between 52 and 50 Ma. The temperature in our VOL simulations does not have exact same trend in temperature as the 10 Myr-averaged temperatures depicted by the oxygen isotopes (Fig. 6). While there is good agreement for the 52 Ma, 30 Ma and 15 Ma time periods, our VOL simulation tends to overestimate the sea surface temperature for the 65 Ma time period. Conversely, the temperature from the UNI simulation fits the oxygen isotope estimates better at the 65 Ma periods, modeled with values of atmospheric CO2 level of 340 ppm. We note here that this estimate corresponds to the atmospheric CO2 level estimated by Beerling and Royer (around 400 ppm on their curve). However, this is below the threshold value for glaciation of Antarctica and thus too low for the 65 Ma time period. Nevertheless, the values of the glaciation thresholds are based on modern and Oligocene continental configurations. At the beginning of the Paleocene however, Australia was still attached to Antarctica. This configuration may be in favor of a lower atmospheric CO2 threshold. With a larger continental mass centered around the pole, warmer summer temperature should induce lower snow accumulation, reducing the atmospheric CO2 threshold for southern continental glaciation. This hypothesis should be tested in a coupled ice-climate model in the future.

The Eocene is a warm period in the model as the result of the intense degassing rate that can largely overcompensate for the elevated CO2 consumption linked to the northward drift of the Deccan basalts. The decrease in degassing rate from the Eocene to the Oligocene causes atmospheric CO2 to cross the threshold of Antarctic glaciation, which might have happened between 52 and
30 Ma. To narrow down when the glaciation of Antarctica started within this time window requires additional continental configurations. The model suggests that the mid-latitude averaged temperature at the Oligocene is close to the pre-industrial value (14 °C) and consistent with the presence of an ice sheet on Antarctica. During the Miocene, the modeled atmospheric CO₂ and temperature increases in response to a decrease of continental weathering. The northward drift of India and Africa tends to reduce the areas of granitic and basaltic lithologies below the ITCZ during that time period. The resulting steady state CO₂ level is 475 ppm, which is consistent with the 400–500 ppm estimation of Kürschner et al. (2008) at the beginning of the Middle Miocene (15.5 Ma). Our result is also consistent with the modeling work by Hamon et al. (2012) but disagrees with estimates from marine phytoplankton carbon isotopes which estimate a pCO₂ as low as 280 ppmv (Pagani et al., 2005, 2011). However, a warmer Miocene is in agreement with the record of the Mid-Miocene Climate Optimum (MMCO) (Kürschner et al., 2008; Mosbrugger et al., 2005; Zachos et al., 2001) which could be related at least in our model to a decrease in the CO₂ consumption rate by weathering.

5. Conclusions

We used a spatially resolved climate-carbon model (GEOCLIM) and investigated the influence of continental drift on the long-term evolution of the Cenozoic atmospheric CO₂. Our model shows that the continental configuration should set the required conditions for the onset of a large-scale glaciation at the beginning of the Cenozoic. This is true even when we imposed in the model a decrease of the weathering owing to the absence of major uplift during the early Cenozoic. The dispersed configuration and the location of a large continental area (North Africa, northern South America) within the ITCZ promoted the CO₂ consumption by weathering, forcing CO₂ to remain low.

Using a 2D representation of the basaltic provinces in the model, we found that Paleocene had higher atmospheric CO₂ because of the location of the Deccan traps in the subtropics, which triggered lower CO₂ consumption by weathering. The weathering effect of this northward drift of the Deccan traps at the Eocene is so high that atmospheric CO₂ is too low in triggering the buildup of the Antarctic ice sheet before the observed pattern. This result is in disagreement with the scenario suggested by Kent and Muttoni (2008). Here we suggest that Antarctic ice sheet did not develop before the Oligocene because of a high degassing rate at the Eocene. In other words, our model results suggest that the transition from a greenhouse Eocene to an icehouse Oligocene is mainly the result of a reduction in CO₂ degassing rate.

Finally during the Miocene period, our model results show that the northward drift of Africa and India causes CO₂ removal to decrease with the weathering, followed by an increase in the atmospheric CO₂ level to up to 500 ppm. The modeling of the
effect of the continental drift on atmospheric CO₂ favors the presence of a Middle Miocene Climate Optimum. Finally, we found that the decrease of the atmospheric CO₂ level from 500 ppm during the Miocene to its pre-industrial values is induced by an increase in silicate weathering, due to a more intense South-East Asian monsoon.
Fig. 6. Evolution of the mid-latitude (32°–58° N and S) sea surface temperature calculated from a carbonate δ18O compilation (Prokoph et al., 2008). Gray squares are the 10 Myr averaged temperature of this dataset, the vertical bars stand for the standard deviation. The blue triangles are the 30–60° S and N sea surface averaged temperature calculated by FOAM accounting for a variable solid Earth degassing (VOL simulation). The vertical bars represent the maximum and minimum calculated temperatures occurring in these areas. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

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Appendix A. Supplementary materials

Supplementary data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.epsl.2013.03.049.

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