Response to anonymous referee #1

First, I would like to thank the referee for his comments and encouragements. His main point concerns the role of petrogenic organic carbon, which represents a significant contributor to the long-term carbon cycle with also a significant role on the isotopic budget.

Indeed, my model considers only three sources and sinks of carbon : volcanic carbon (V) which is always a source ; carbonate precipitation (D) which always represents a net sink, though both dissolution and accumulation are considered through carbonate compensation; and finally organic carbon (B) which corresponds both to sinks (burial of recent organic matter), but also possibly to sources (oxidation of old organic carbon). Though this was not explicitly detailed in the manuscript, this last possibility (ie. a negative contribution to B, or "negative" burial) can be in part interpreted as a petrogenic organic carbon source. So implicitly, the model does already include petrogenic organic carbon. But, as explained by reviewer #1, this point needs to be discussed more precisely in a revised manuscript, since the negative contributions to B were only described as « net "old" soil erosion and remineralization » in the original submitted paper. Clearly, this was misleading.

As summarized by reviewer #2, the model is based first of all, on a rather standard steadystate equation for carbon. From the isotopic balance equation (2b), we deduce that the baseline (long-term) value B_0 for all organic fluxes, including petrogenic ones, should be about 20% of the volcanic flux, that is $B_0 = V/5$, in order to account for observed isotopic compositions. This baseline value B_0 represents the sum of positive terms, mostly due to the burial of recent organic matter, but also negative ones that correspond to the oxidation of "old" soils and indeed "petrogenic" or "fossil" organic matter. As underlined by reviewer #1, the absolute magnitude of each term is currently not well constrained and positive and negative contributions to B_0 are, individually, possibly comparable to V: indeed, if V is taken in the range of 40 to 175 TgC/yr (Burton et al., 2013), the estimate for petrogenic organic carbon from Blair et al. (2003) [36 to 48 TgC/yr] corresponds to the lower range of V. This certainly needs to be explained in the revised manuscript.

=> I have now added the following paragraph in the revised manuscript:

(lines 107-113)

"It must be stressed that B stands for all organic carbon fluxes, whether they correspond to organic carbon burial (positive contributions to B) or to organic matter oxidation (negative contributions to B). If the long-term average equilibrium value B_0 needs to be positive to account for the isotopic balance as shown above, this is not necessary always the case for the instantaneous values of B, as we will illustrate it below with the astronomical forcing. Indeed, B represents a sum of positive and negative terms whose individual absolute magnitudes are much larger than the long-term net value B_0 . For instance, the oxidation of petrogenic organic carbon alone will contribute negatively to B, with a magnitude that may be as large as 40 TgC/yr (Blair et al., 2003)."

=> and also the mention of petrogenic organic carbon in the conclusion:

(lines 282-283)

"This model was built on the premises that changes in organic matter or petrogenic organic carbon fluxes are responsible for the 400-kyr oscillations observed in Cenozoïc ¹³C records"

Still the main point of the paper was not about the detailed steady state balance of the carbon system, but about its possible dynamics over the last 4 million years. For my model equations, only the net values of B₀ (or B) are relevant. As explained by reviewer #1, the dynamics of petrogenic organic matter fluxes will depend on erosion and continental runoff. It will therefore contribute to the generic situation described in the manuscript, or "Amazon-like" situation, with enhanced organic carbon oxidation when precession maxima favours more precipitation and erosion. More precisely, when including precessional forcing through B = B₀ - *a* F(t) with the numerical values B₀ = 25 TgC/yr , *a* = 50 TgC/yr (see legend of Fig.2), then the "net burial" B does change sign through time, and becomes temporarily a carbon source when negative: it is then dominated by the oxidation of organic matter (soil, but also fossil or petrogenic...).

=> I also have added the following sentence:

(lines 146-148) "But this recent soil together with older soils and with petrogenic organic carbon (Galy et al., 2008) will be eroded and transported to the ocean through enhanced river discharges."

Response to anonymous referee #2

First, I would like to thank the referee for his comments and support. He addresses several important technical points listed below and makes a more general remark, that my conceptual model is rather generic and could correspond to other geomorphological mechanisms than the one described in the manuscript. I believe that most of his comments can easily be addressed by a more explicit description of the model, its parameters, and its results, as explained below on a point-by-point basis.

RC1: For component (1), I see no error in the carbon cycle equations as written, but there are a few steps/assumptions that are not clearly articulated. Adding more details deriving each equation would make the paper easier to follow. In equation (1a) it is implicitly assumed that the weathering and volcanic fluxes can be lumped together (which is fine based on the assumption that both approximate the mantle isotopic value), though this is not stated. (Otherwise the equation should be dC/dt = V + W - B - D).

AC1 : I somewhat disagree on this point. Silicate weathering W takes one CO_2 molecule from the atmosphere (or more precisely H_2CO_3 from precipitation and runoff) and transforms it into HCO_3^- (through acido-basic reaction or proton exchange) in the river system and finally the ocean. When considering the oceanic carbon budget alone, W indeed adds one carbon in the ocean. But I am considering the "global" Earth surface budget (ocean + atmosphere) and therefore W has no net effect on C. Therefore W does not appear in equation (1a) for dC/dt. Its impact on the global carbon cycle arises only through the ocean alkalinity budget (dA/dt) and carbonate compensation, which leads to carbonate deposition D being directly linked to silicate weathering through D = W-V+B.

=> I now insist on the fact that C includes both the ocean and atmosphere, and better explain the underlying mechanisms.

(lines 91-95)

Silicate weathering W takes one CO_2 molecule from the atmosphere, or more precisely one H_2CO_3 from precipitation and runoff, and transforms it into a HCO_3^- that finally reach the ocean. When considering the "global" Earth surface budget C which includes the ocean and atmosphere, W has therefore no direct effect on C and does not appear in equation (1a) for dC/dt, but only as a source of alkalinity in equation (1b).

RC2 : Next, I think it would be helpful to start with the full version of equation (2b): $d/dt(\delta C^*C) = V^*\delta V - B^*\delta B - D^*\delta D$

Then it would be more straightforward to see how the final version is obtained through the product rule and assumption that $\delta c = \delta D$ as well as constant values of $\delta V = -5\%$ and $\delta B = -25\%$. This is particularly important because it is more typical to describe a constant fractionation of organic carbon with respect to δC , rather than a constant δB .

AC2 : This is indeed a good idea. This corresponds also to the remark from Peter Köhler (doi:10.5194/cp-2017-3-SC1) that the equations should be clarified, and the underlying assumptions should be more explicit.

=> I now add the derivation of equation (2b) and explicit choices for $\delta^{13}V$, $\delta^{13}B$, $\delta^{13}D$.

(lines 114-125) The isotopic ¹³C budget can be written as: $d/dt(C\,\delta^{13}C) = V\,\delta^{13}V - B\,\delta^{13}B - D\,\delta^{13}D$ where $\delta^{13}C$ is the isotopic composition of ocean carbon, $\delta^{13}V$ the isotopic composition of the volcanic carbon input, $\delta^{13}B$ the isotopic composition of organic matter and $\delta^{13}D$ the isotopic composition of marine carbonates. This can be re-written as: $C \left(d\delta^{13}C/dt \right) + \left(dC/dt \right) \delta^{13}C = V \,\delta^{13}V - B \,\delta^{13}B - D \,\delta^{13}D$ $C (d\delta^{13}C/dt) = V \delta^{13}V - B \delta^{13}B - D \delta^{13}D - (V - B - D)\delta^{13}C$ or $= V (\delta^{13}V - \delta^{13}C) - B (\delta^{13}B - \delta^{13}C) - D (\delta^{13}D - \delta^{13}C)$ If we neglect isotopic fractionation during carbonate precipitation (in other words, $\delta^{13}D$ = $\delta^{13}C$) and more generally during carbonate compensation, we finally obtain: (2b) $d\delta^{13}C/dt = (V(\delta^{13}V \cdot \delta^{13}C) - B(\delta^{13}B \cdot \delta^{13}C))/C$ In the following we will assume a constant -5% volcanic source $\delta^{13}V$, as well as a constant -25% organic matter value $\delta^{13}B$ (eq. Porcelli and Turekian, 2010).

RC3: On that note, adding an appropriate subscript to the δ notation (rather than writing as δ^{13}) would be helpful to differentiate between the δ values for each flux.

AC3 : I will follow this suggestion and write the final equation (2b) as:

 $d\delta^{13}C/dt = (V(\delta^{13}V - \delta^{13}C) - B(\delta^{13}B - \delta^{13}C))/C$

=> Equation (2b) is now written as: (line 123) (2b) $d\delta^{13}C/dt = (V(\delta^{13}V-\delta^{13}C) - B(\delta^{13}B-\delta^{13}C))/C$

RC4: Finally, there should be explanation of scaling between pCO2 and total C (namely, that the assumptions are being made that the ocean inventory of Ca2+ does not change and that the mass of carbon in the system is well-approximated by the ocean bicarbonate pool).

AC4 : This corresponds also to the remark from Peter Köhler. As explained in my response (doi:10.5194/cp-2017-3-AC1), this will be justified in the revised version.

=> The scaling of pCO2 is now explained in the text as followed:

(lines 126-130)

Indeed, if we assume, to first order, that C may represent the carbon content of a wellmixed ocean, then from chemical equilibrium pCO_2 should be proportional to $[HCO_3^-]^2/$ $[CO_3^{2-}]$. After carbonate compensation (ie. assuming that $[CO_3^{2-}]$ remains constant) and considering that C is dominated by bicarbonates $[HCO_3^-]$ under standard pH conditions, we end up with the approximate scaling that pCO_2 varies roughly as C^2

(lines 94-96)

if we assume that the oceanic calcium concentration does not change significantly over the last few millions of years, carbonate compensation will restore the oceanic carbonate content. RC5 : For component (2), it would be helpful to provide the chosen value for the scaling term a in equation (3) in the text and not just the caption to Fig. 2. Later in the paper, it is mentioned that a has to be of the same order as the equilibrium organic C burial flux, but the value in the caption is in fact double the equilibrium burial flux. There should also be a description of how this value was determined (presumably to get the right amplitude in the modeled δc)?

AC5 : I agree that a better discussion of parameter values could be included in the text, though these values are indeed determined empirically in order to get a qualitatively correct response. The amplitude *a* is indeed the double of the equilibrium flux B_0 for the particular experiments shown on Fig.2. The comment in the text was slightly more generic (*« the strength of the forcing a needs to be of the same order than the baseline value B*₀. *This is a robust feature, which does not depend on model setting or parameters »*).

I will add a short discussion on the choices made for *a*. I will rewrite the above sentence somewhat differently, as *«when variations in B (or equivalently the parameter a) are smaller than its baseline value B₀, the model cannot reproduce the amplitude of \delta^{13}C observed in marine benthic records ».*

 \Rightarrow I have now added a short discussion on the choices made for *a*, and modified the ambiguous sentence.

(lines 161-162) The value of the parameter a is chosen in order to obtain approximately the correct amplitude for these simulated 400-kyr oscillations (a = 50 TgC/yr).

(lines 217-218) When variations in *B*, as determined by parameter *a*, are smaller than the baseline value B_0 , the model cannot reproduce the oceanic amplitude of $\delta^{13}C$ observed in marine benthic records

RC6 : To me, component (3) is the most novel element of this conceptual model. This threshold term allows for a switch between two styles of periodic forcing of the organic carbon burial flux. In general, the periodic forcing reduces the value of B, except if the sedimentary reservoir is near to its maximum size, in which case periodic forcing switches to increasing the value of B...Next, what is the basis for setting the threshold condition at S < 0.85 SMAX? The text notes that this threshold mechanism causes a switch in organic carbon burial after significant sea level drops at 2.4-2.5 Myr and 0.35-0.65 Myr, but was the threshold set in order to provide this result?

AC6 : I should certainly also be more explicit here. The "normal" (pre-Quaternary) situation ("progradation") is indeed when the periodic forcing reduces the value of B. Then the sedimentary reservoir S is typically at its maximum (we have S=Smax) as shown on Fig.2. But after every significant new sea-level drop (from the zmin "river incision" curve based on Lisiecki and Raymo, 2005), Smax = $zmin^3$ increases significantly and the situation is switched to "aggradation". This first switch ("switch ON") is not strongly dependent of the 0.85 threshold parameter, since a sea-level drop as small as about 5% will induce a sufficient increase in Smax (=15%) to trigger the change. But the switch back to

normal ("switch OFF") and therefore the duration of the "aggradation" phase, will depend more strongly on this threshold choice. In other words, concerning the two major "aggradation" phase discussed in the text (2.4-2.5 Myr and 0.35-0.65 Myr), their starts are directly linked to the significant sea level drops (at 2.5 Myr and 0.65 Myr): they are independent of the threshold value. But their duration is rather directly linked to this threshold value of 0.85 and also to the choice of parameter b. The choice of a different threshold value than 0.85 will consequently affect the amplitude of the differences between experiments (b) and (c), but not the timing of these differences.

=> This is now explained in the revised manuscript.

(lines 193-196)

After a short transient period, this reservoir remains therefore equal to this maximum value S_{MAX} in the absence of major sea level drops, as during the pre-Quaternary period. In contrast, for each significant sea level drop, S_{MAX} increases abruptly and we start a new transient phase whose duration is linked to parameter b.

(lines 206-207)

The start of these transient periods is directly linked to sea level drops, according to the LR04 forcing, while the duration of these transients is linked both to the 0.85 S_{MAX} threshold and the b parameter.

RC7 : Again, the value of the scaling factor for the growth rate of the sedimentary reservoir, *b*, should be provided in the text, along with an explanation of how this value was determined.

AC7 : I agree. And again, the value of b is a rather empirical choice. As explained above, its value will affect the duration of "aggradation" phases, and consequently the amplitude of the differences between experiments (b) and (c).

 \Rightarrow I now present the rational behind choices made for *b*.

(lines 207-208)

... whose values are chosen to qualitatively better match the $\delta^{13}C$ data. For the results show on Fig.2, $b = (160 \text{ kyr})^{-1}$.

RC8 : Also, in Figure 2, it is clear to see why the addition of this threshold term appreciably changes model behavior around 0.6 Myr, but not obviously earlier in the record. Maybe this is just hard to see because of the scale on the axes?

AC8 : There is indeed a significant change around 0.6 Myr that explains why the "400 kyr ¹³C cycles" are disturbed at this time. There is also a significant change at about 2.4 MyrBP in the ¹³C results on Fig.2 (experiment (c): red curve) whereas the results without this mechanism (experiment (b): blue curve) the simulated ¹³C values are significantly out of the range of observed values. Interestingly, the "switch" model was designed to address the disturbed "400 kyr ¹³C cycles" of the last 1 MyrBP. The better agreement with data at 2.4 MyrBP was not expected, and comes as a bonus.

=> I now clarify the role of the threshold mechanism when discussing results shown on Fig.2, and I have added a short comment on this last point in the discussion.

(lines 210-211) "...as illustrated by the difference between the blue and red curves on Fig.2".

(lines 250-251)

This mechanism also allows for simulated marine $\delta^{13}C$ in better agreement with data at about 2.4 MyrBP.

=> but this comment also probably arises because, if the mode-switch is rather clear on Fig.2 when looking at the ¹³C results, it is less so on the geomorphological variables s and Smax. I have therefore added an orange shading on top of these curves, corresponding to the transient aggradation regimes (S < 0.85 S_{MAX}).

(added shading in Fig. 2)

RC9 : However, it does not seem that the conceptual model is particularly linked to the mechanism proposed (a shift between progradational to aggradational river systems). Paillard suggests in the introduction that "astronomical parameters are influencing climate through other mechanisms than the growth and decay of ice sheets," but it seems to me that what's been done is to link organic carbon burial to the growth and decay of ice sheets via the impact on sea level. This means the conceptual model is equally applicable to any process related to sea level that can drive a threshold response in organic carbon burial. This is not a flaw in the conceptual model, but parts of the text could be rewritten to emphasize that the geomorphological mechanism is only one possible physical interpretation of what the model actually describes.

AC9 : The first aim of this model is to link the observed 400-kyr ¹³C oscillations and the associated carbon cycle changes to the astronomical forcing, through the dynamics of organic matter burial. This is in general fully independent of sea level changes, except for the most recent Quaternary period. Since our knowledge of the carbon cycle is much more detailed over this recent period (pCO₂ data, numerous ¹³C records, ...), it is necessary to explain both the rather generic 400-kyr ¹³C oscillations observed during the Cenozoïc and beyond, but also why the Quaternary ¹³C oscillation look different and how this relates to observed pCO₂ fluctuations. As explained in the introductory part of the paper, I am using a deductive line of thought. I certainly agree with the reviewer that the mechanism suggested here is probably not the only possible one. It is nevertheless (to my knowledge) the first one suggested so far that may explain both the recent past and the more remote one, in the same conceptual framework.

=> I have added the following paragraph in the conclusion

(lines 282-288)

"This model was built on the premises that changes in organic matter or petrogenic organic carbon fluxes are responsible for the 400-kyr oscillations observed in Cenozoïc ¹³C records, and that the large sea-level variations occurring during the Quaternary are strongly affecting this process. Continental margins and sedimentary fans are a very likely key component, as illustrated by our simple conceptual model. But obviously, many complex processes are involved in the interactions between organic matter burial or oxidation, monsoons and sea-level changes. The geomorphological mechanism described here is one possibility which allows, for the first time, to account both for the persistent 400-kyr oscillation observed in ¹³C records during the Cenozoïc, but also for its change during the last million years"

RC10 : Also, more discussion about the relationship between pCO2 and $\delta 13C$ cycles represented by this conceptual model would be welcome. Based on the introduction, I expected further explanation of phasing between simulated cycles and eccentricity. In particular, how well has the model accounted for a change in the nature of the 400 kyr $\delta^{13}C$ oscillation in the last million years?

AC10 : Indeed, it is probably important in the discussion to re-state the main objective of this model: reproducing not only the 400 kyr δ^{13} C oscillation seen during the pre-Quaternary, but also explaining why it is perturbed during the last million years, and to insist on the final δ^{13} C conclusion: assuming that this perturbation is caused by major sea level drops, as performed in this model, leads not only to a better agreement for the δ^{13} C curves, but also explains several features of the CO₂ changes.

=> This is now discussed in more details in the discussion,

(lines 240-251)

"According to this mechanism, in the ordinary sedimentary regime (progradation), we obtain changes in the carbon cycle with pCO_2 maxima and $\delta^{13}C$ minima associated directly to eccentricity maxima. This is indeed consistent with long Cenozoïc records (eg. Pälike et al, 2006).

When we allow for changes in the sedimentary regime triggered by sea level changes, the model can also reproduce more peculiar features. Indeed, up to now it has been difficult to explain the last two long-term cycles observed in the marine $\delta^{13}C$, each being approximately 500 kyr-long, with a maximum now (δ^{13} Cmax-I), a well-marked maximum at about 500 kyr BP (δ^{13} Cmax-II) and a previous one around 1000 or 1100 kyr BP (δ^{13} Cmax-II). In the model described here, these two long oscillations are generated from the eccentricity forcing, but with an abrupt switch to aggradation mode at about 620 kyrBP caused by the sea level drop at MIS 16. This switch reverses the phase of the 400-kyr carbon oscillation during a few hundred thousands of years. Interestingly, this also induces a slight minimum in the carbon (or pCO₂) results, consistent with the observed low pCO₂ values observed in the Antarctic ice core around 600-700 kyrBP. This mechanism also allows for simulated marine $\delta^{13}C$ in better agreement with data at about 2.4 MyrBP. »

RC11 : Also, why is the 100 kyr term added only to the modeled $\delta 13C$ and not pCO2?

AC11 : The 100-kyr term added to the ¹³C results (orange curve) is just an "ad-hoc" addition to improve the match with data, based on the (usually accepted) hypothesis that this 100-kyr oscillation in the ¹³C is linked to the varying size of the biosphere. There is no such data for the pCO₂ over the last 4 million years, and there is no simple explanation for the observed pCO₂ 100-kyr cycles: adding this cycle *a posteriori* is therefore certainly not justified for pCO₂. More importantly, the 100-kyr cycle is not the subject of this manuscript, so may be I should simply remove the orange curve to simplify the figure.

=> I have removed the 100-kyr "ad-hoc" addition on the ¹³C results (orange curve) and the corresponding paragraph in the main text.

(removed lines 215...)

RC12: Perhaps add the eccentricity and filtered eccentricity to the same figure as the modeled curves.

AC12 : Yes. This would indeed simplify the discussion of the results in terms of phasing, according to the above comments (RC10).

=> I have added the filtered eccentricity at the bottom of Fig.2, as well as the filtered ${}^{13}C$ data and model results. This new figure setting corresponds also better to the data shown on Fig.1, in order to simplify the understanding of the figure.

(added curves in Fig. 2)

RC13 : Finally, in the results section of the text, comparison between blue and black curves in Figure 2 is cited as evidence for good agreement between model results and observations, but both these curves are model results.

AC13: This was a bad formulation in the original text. I meant that experiments a and b, with and without the long-term trend, (ie. the black and blue curves) were very similar in terms of ¹³C, and both were comparable to the data (the grey curves).

=> This is now changed in the text as:

(lines 163-165)

More specifically, the $\delta^{13}C$ black and blue simulated curves are superimposed and almost undistinguishable, since the linear trend added to the carbon cycle has almost no impact on the $\delta^{13}C$. They are both most of the time within the range of observed values (gray curves).

Response to Peter Köhler

First, I would like to thank Peter Köhler for providing these thoughtful comments and for rebuilding and reproducing my model results. I acknowledge that the model description and the presentation of some equations or parameters was sometimes not explicit enough in the submitted manuscript. I therefore want to clarify some points below.

1. The ¹³C equation.

The main point raised by Peter Köhler concerns the 13 C equation. It turns out that we are both using (almost) the same equation, but just written differently. The only true difference stands in the organic matter fractionation : while Peter Köhler uses a constant fractionation with respect to the environment, I implicitly considered an organic matter sink with a constant isotopic signature of -25‰.

More precisely, equation (8) from Peter Köhler reads :

$$\frac{d}{dt}(\delta^{13}C) = \frac{1}{C} \left[V(-5) - B(\delta^{13}C - 25) - (W + B - V)\delta^{13}C - \delta^{13}C \cdot \frac{dC}{dt} \right]$$

When substituting the last term using the equation for dC/dt :

$$\frac{d}{dt}C = V - B - D$$

we get :

$$\frac{d}{dt}(\delta^{13}C) = \frac{1}{C}[V(-5) - B(\delta^{13}C - 25) - (W + B - V)\delta^{13}C - \delta^{13}C.(V - B - D)]$$

And after simplification, and using D = W+B-V, this leads to :

$$\frac{d}{dt}(\delta^{13}C) = \frac{1}{C}[V(-5 - \delta^{13}C) - B(-25)]$$

As mentioned above, this is very similar to my equation (2b) :

$$\frac{d}{dt}(\delta^{13}C) = \frac{1}{C}[V(-5 - \delta^{13}C) - B(-25 - \delta^{13}C)]$$

the only difference being that, implicitly, I used a constant organic matter sink of -25‰. Since the ¹³C of carbonates remains close to 0‰, these different choices lead to a very small difference in the numerical experiments, as demonstrated by Peter Köhler. In any case, this point should be clarified in a revised manuscript.

=> Following also the advice of Rev.#2 (RC2), I now add the explicit derivation of equation (2b) and explicit choices for $\delta^{13}V$, $\delta^{13}B$, $\delta^{13}D$.

(lines 114-125) The isotopic ¹³C budget can be written as: $d/dt(C \ \delta^{13}C) = V \ \delta^{13}V - B \ \delta^{13}B - D \ \delta^{13}D$ where $\delta^{13}C$ is the isotopic composition of ocean carbon, $\delta^{13}V$ the isotopic composition of the volcanic carbon input, $\delta^{13}B$ the isotopic composition of organic matter and $\delta^{13}D$ the isotopic composition of marine carbonates. This can be re-written as: $C (d\delta^{13}C/dt) + (dC/dt) \ \delta^{13}C = V \ \delta^{13}V - B \ \delta^{13}B - D \ \delta^{13}D$ or $C(d\delta^{13}C/dt) = V \delta^{13}V - B \delta^{13}B - D \delta^{13}D - (V - B - D)\delta^{13}C$ $= V(\delta^{13}V - \delta^{13}C) - B(\delta^{13}B - \delta^{13}C) - D(\delta^{13}D - \delta^{13}C)$ If we neglect isotopic fractionation during carbonate precipitation (in other words, $\delta^{13}D$ $= \delta^{13}C$) and more generally during carbonate compensation, we finally obtain: (2b) $d\delta^{13}C/dt = (V(\delta^{13}V - \delta^{13}C) - B(\delta^{13}B - \delta^{13}C))/C$ In the following we will assume a constant -5‰ volcanic source $\delta^{13}V$, as well as a constant -25‰ organic matter value $\delta^{13}B$ (eg. Porcelli and Turekian, 2010).

2. The pCO₂ scaling equation.

In the manuscript, C represent the total carbon content at the « Earth surface », which means mostly the ocean reservoir, plus a minor contribution from the biosphere and atmosphere. I used a simple scaling to translate these changes in carbon content C (expressed in GtC) in terms of atmospheric pCO_2 (in ppm) :

$$pCO_2 = 280 \left(\frac{C}{40,000}\right)^2$$

As explained by Peter Köhler, this might be supported by model experiments for long time scales, but this lacks some justification in the manuscript. Such a scaling can be obtained when considering that C represent the carbon content of a well-mixed ocean. Then, from chemical equilibrium, we obtain :

$$pCO_2 = k \frac{[HCO_3^-]^2}{[CO_3^{2-}]}$$

where the constant *k* includes the solubility of CO₂, and the first and second dissociation constants of carbonate and bicarbonate ions. When considering only the long time scale response, we can assume that carbonate compensation will restore $[CO_3^{2-}]$ to a constant initial value. Furthermore, under standard oceanic pH conditions, bicarbonate ions $[HCO_3^{-}]$ represent about 90% of the total carbon content C. If we assume, to first order, that $C \approx [HCO_3^{-}]$, then the above equation means that pCO₂ should, on long time scales, increase approximately as the square of C. Though this is certainly a rough approximation, it is sufficient to provide a reasonable magnitude of the implied pCO₂ changes associated with this simple model.

Again, this point should be clarified in a revised manuscript

=> Following also the advice of Rev.#2 (RC4), the scaling of pCO2 is now explained in the text as followed:

(lines 126-130) Indeed, if we assume, to first order, that C may represent the carbon content of a wellmixed ocean, then from chemical equilibrium pCO_2 should be proportional to $[HCO_3^-]^2/$ $[CO_3^{2-}]$. After carbonate compensation (ie. assuming that $[CO_3^{2-}]$ remains constant) and considering that C is dominated by bicarbonates $[HCO_3^-]$ under standard pH conditions, we end up with the approximate scaling that pCO_2 varies roughly as C^2

(lines 94-96)

if we assume that the oceanic calcium concentration does not change significantly over the last few millions of years, carbonate compensation will restore the oceanic carbonate content. 3. I indeed also used the rather implicit assumptions that ocean alkalinity is approximated by carbonate alkalinity, therefore equation (1b) in the manuscript. This could be discussed a bit more in the manuscript, though it is quite a classical approximation.

=> addition (line 90) ...assuming that alkalinity is dominated by carbonate alkalinity.

Concerning the choice of precessional forcing $F_0(t) = \max(0, -e \sin\omega)$, I am not sure that any proxy comparison would either backup or dismiss such a choice. Furthermore, there is little hope to find any proxy for global organic carbon preservation, since individual proxies of preservation are often very dependent of the local or regional context. The choice of this forcing is simply based on two premises: 1 – monsoon are primarily driven by precession, something demonstrated by paleoclimatic data and simulated by climate model. 2 – the conceptual model needs a rectifying mechanism to reproduce the envelope of precession, something consistent with the averaged values of river sedimentary carbon discharges being largely dominated by the largest or extreme events. The expression above is the simplest possible choice along these lines.

The isotopic signatures used for volcanic outgassing (-5‰) and for buried organic matter (-25‰) are rather standard values used in geochemical textbooks and treatises. For instance:

Porcelli, D. and Turekian, K.K., The History of Planetary Degassing as Recorded by Noble Gases, §6.6.1 in Readings from the Treatise on Geochemistry, edited by Holland, H.D. and Turekian, K.K., (2010).

These numbers are somewhat conventional with actual measurements varying from about -1‰ to -8‰ for volcanoes or mid-ocean ridges outgassing, depending on location. Similarly, -25‰ is a conventional value for organic matter δ^{13} C used for instance as a normalization for reporting ¹⁴C activities, while actual values vary from roughly -10‰ to -30‰ depending on organic materials.

=> reference added

(lines 124-125) In the following we will assume a constant -5% volcanic source δ^{13} V, as well as a constant -25% organic matter value δ^{13} B (eg. Porcelli and Turekian, 2010).

4. There is no need to specify the carbon content of the model, since it is explicitly computed by the equations. As mentioned in Figure caption 2, the model is integrated from an arbitrary condition (that is carbon content, and isotopic value) at 5 MyrBP and the first 1 Myr is discarded, since it correspond to the transient part of the simulation.

=> This is now also explicitly mentioned in the text.

(lines 133-134) The model is integrated from an arbitrary initial condition at 5 MyrBP and the first 1 Myr is discarded. 5. I believe all parameter values are given in Figure caption 2, but there has been an unfortunate typesetting change from greek to latin alphabet. This also needs to be corrected in a revised manuscript.

=> typesetting has been checked => parameter values are now given also in the main text (cf. Rev.#2, RC5 & RC7). (line 161) (a = 50 TgC/yr). (line 207) b = (160 kyr)⁻¹. (line 132) γ set to 1,2 TgC/yr

The Plio-Pleistocene climatic evolution as a consequence of orbital forcing on the carbon cycle

(revised, August 9. 2017)

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Abstract. Since the discovery of ice ages in the XIXth century, a central question of climate science has been to understand the respective role of the astronomical forcing and of greenhouse gases, in particular changes in the atmospheric concentration of carbon dioxide. Glacial-interglacial cycles have been shown to be paced by the astronomy with a dominant periodicity of 100 ka over the last million years, and a periodicity of 41 ka between roughly 1 and 3 million years before present (MyrBP). But the role and dynamics of the carbon cycle over the last 4 million years remain poorly understood. In particular, the transition into the Pleistocene about 2.8 MyrBP or the transition towards larger glaciations about 0.8 MyrBP (sometimes refered as the mid-pleistocene transition, or MPT) are not easily explained as direct consequences of the

- 15 astronomical forcing. Some recent atmospheric CO₂ reconstructions suggest slightly higher pCO₂ levels before 1 MyrBP and a slow decrease over the last few million years (Bartoli et al., 2011; Seki et al., 2010). But the dynamics and the climatic role of the carbon cycle during the Plio-Pleistocene period remain unclear. Interestingly, the δ^{13} C marine records provide some critical information on the evolution of sources and sinks of carbon. In particular, a clear 400-kyr oscillation has been found at many different time periods and appears to be a robust feature of the carbon cycle throughout at least the last 100 Myr (eg.
- 20 Paillard and Donnadieu, 2014). This oscillation is also visible over the last 4 Myr but its relationship with the eccentricity appears less obvious, with the occurrence of longer cycles at the end of the record, and a periodicity which therefore appears shifted towards 500-kyr (cf. Wang et al., 2004). In the following we present a simple dynamical model that provides an explanation for these carbon cycle variations, and how they relate to the climatic evolution over the last 4 Myr. It also gives an explanation for the lowest pCO₂ values observed in the Antarctic ice core around 600-700 kyrBP. More generally, the
- 25 model predicts a two-step decrease in pCO_2 levels associated with the 2.4 Myr modulation of the eccentricity forcing. These two steps occur respectively at the Plio-Pleistocene transition and at the MPT, which strongly suggests that these transitions are astronomically forced through the dynamics of the carbon cycle.

1 Introduction

The idea that the orbital parameters of the Earth may influence climate has a long history, linked mostly to the development

- 30 of theories of ice ages (eg. Paillard, 2015). But it is quite clear from geological records, that astronomical climatic variations are occuring throughout the Earth history, with or without ice being present on Earth. It is therefore certain that the astronomical parameters are influencing climate through other mechanisms than the growth and decay of ice sheets. This is for instance well-known concerning records of monsoons or records of low latitude precipitations, which are strongly influenced by precession. A very illustrative example is given by the Mediterranean sapropels (Lourens et al., 1996; Hilgen
- 35 et al., 1999) which are used to calibrate the 40 Ar/ 39 Ar radiochronometers (Kuiper et al, 2008). Similarly, a 400-kyr oscillation is observed in the δ^{13} C of the foraminifera recovered from marine records, throughout most of the Cenozoic (Pälike et al., 2006; Cramer et al., 2003; Sexton et al., 2011; Billups et al., 2004; Wang et al., 2010). This oscillation is present in the benthic records, but also in many planktic ones, suggesting that this δ^{13} C variations are linked to global ocean δ^{13} C changes. This persistent oscillation was recently used to reconstruct the evolution of the Earth's carbon over the last 100-Myr (Paillard
- 40 and Donnadieu, 2014). A key difficulty is to understand the dynamics of this cycle. In particular during the last million years, these oscillations appear to stretch and the relationship with eccentricity becomes less clear (eg. Wang et al., 2004; 2010), as illustrated on Figure 1.

Before 1 MyrBP when ice sheets remained medium sized, the cyclicity appears locked to eccentricity, with high eccentricity values associated with decreasing or low values in $d^{13}C$. This phase relationship appears consistent with earlier time periods,

- 45 with the chronology of Cenozoïc marine cores being sometimes based on the association of high eccentricity and low δ^{13} C values (eg. Paelike et al., 2006; Cramer et al., 2003). A simple deduction is that, most probably, the dynamics behind this oscillation is essentially stable and linked to the astronomical forcing before 1 MyrBP, but it is strongly disturbed by the large Quaternary glaciations afterwards. This observation has major implications on the possible mechanisms, as we will see further on.
- 50 There is no consensus on the cause of these δ^{13} C oscillations, but monsoons or the associated low latitude precipitations are known to respond to precessional forcing, and therefore to be modulated by the 400-ky eccentricity cycles. Still many factors may contribute to the evolution of the carbon cycle on these time scales, like erosion, vegetation dynamics, ocean biogeochemical or dynamical changes. It was therefore suggested that the δ^{13} C cycles could be caused by the modulation of weathering in monsoonal regions (Paelike et al 2006) or by ecological shifts in marine organisms, possibly linked to nutrient
- 55 availability (Wang et al, 2004; Rickaby et al, 2007). It is worth emphasizing that during the last million years, if the link with eccentricity is less obvious, there are clear indications that these δ^{13} C shifts are associated to major changes in the Earth carbon cycle. For instance, carbonate deposition exhibits major changes, well correlated with these global δ^{13} C changes (Bassinot et al, 1994; Wang et al, 2004), and the record of atmospheric pCO₂ from Antarctic ice cores also shows a 10 to 20 ppm long term modulation with a minimum level around 0.6-0.7 MyrBP and a maximum around 0.3-0.4 MyrBP (Lüthi et al,

- 60 2008) in phase with the long term carbonate preservation cycle. A mechanistic modeling of these 400-ky to 500-ky cycles is therefore a critical missing element in our understanding of climate-carbon evolution over the Plio-Pleistocene period. With a simple ocean box model (Russon et al, 2010), it was shown that silicate weathering alone could not account for the simultaneously observed rather large δ¹³C changes (> 0.4‰) and rather small pCO₂ variations (< 20 ppm) in this frequency band during the last million years. Furthermore, with silicate weathering only, the model-predicted phase relationships were</p>
- 65 also inconsistent with observations of δ¹³C, carbonate deposition and pCO₂. Changes in organic matter fluxes are therefore a necessary ingredient in order to account for the observed rather large δ¹³C changes. A possible mechanism could therefore be linked to ocean organic matter burial, associated to changes in nutrient supply or ecological shifts (Rickaby et al, 2007). But it is then very difficult to explain why this mechanism would change drastically with the occurrence of major glaciations, as suggested on Fig. 1. We will therefore build our model on a different perspective, involving a more direct link between 70 monsoons and organic matter burial, that should be strongly affected by sea level changes.
- Organic matter burial takes place mostly on the continental shelves. Recent re-assessments of riverine carbon fluxes to the ocean have emphasized the role of the erosion of continental organic carbon in the overall balance (eg. Galy et al., 2007; Hilton et al., 2015). When investigating the influence of monsoons on the carbon cycle, it is natural to have a closer look at river discharges in monsoonal areas. Carbon budgets on major present-day erosional systems have provided some contrasted
- 75 results, with riverine organic matter being either a net carbon source for the ocean (Burdige, 2005), or a net sink through organic carbon burial in sedimentary fans (Galy et al., 2007). The first study was mostly based on the Amazon basin, while the second estimation is from the Himalayan system. The differences are likely linked to different river basin configurations and different sedimentary deposition dynamics. This dramatically highlights the impact of geomorphology on terrestrial organic carbon burial, and suggests that the long term global balance might be different in a context of large glacial-
- 80 interglacial sea level variations like the last million years, when compared to earlier periods with much smaller sea level changes. Our conceptual model is therefore built on the impact of monsoon-driven terrestrial organic matter burial on the global carbon cycle.

2 Conceptual Model

We are interested by the evolution of the global Earth carbon, that is the carbon content of the atmosphere, the ocean and the biosphere, which amounts today approximately to C = 40,000 PgC (petagrams of carbon, ie. 10^{15}gC). This evolution results from possible imbalances between the volcanic inputs V, the oceanic carbonate deposition flux D associated to silicate weathering and its alkalinity flux to the ocean W, and the organic carbon burial B. Our model equations are:

> (1a) dC/dt = V - B - D(1b) dA/dt = W - 2D

90 where the second equation represents the alkalinity balance, assuming that alkalinity is dominated by carbonate alkalinity. Silicate weathering W takes one CO₂ molecule from the atmosphere, or more precisely one H₂CO₃ from precipitation and

Didier Paillard 8/8/y 11:59 Commentaire [1]: Peter Köhler's comment #3



runoff, and transforms it into a HCO₃⁻ that finally reach the ocean. When considering the "global" Earth surface budget C which includes the ocean and atmosphere, W has therefore no direct effect on C and does not appear in equation (1a) for dC/dt, but only as a source of alkalinity in equation (1b). On time scales larger than several millennia, if we assume that the

95 oceanic calcium concentration does not change significantly over the last few millions of years, carbonate compensation will restore the oceanic carbonate content. Therefore, to first order, we can write :

$$d[CO_3^{2-}]/dt = d(A-C)/dt = 0 = W - D - V + B.$$

Solving for D, this leads to the long-term evolution equations for carbon:

 $(2a) \qquad \qquad dC/dt = 2(V-B) - W$

For simplicity, we will assume that the main stabilizer of the carbon system is the silicate weathering, with a fixed relaxation time τ_C , ie. W = C/ τ_C . Solving the present-day equilibrium with $\delta^{13}_{Eq} = 0\%$ as a typical value for carbonates, we easily deduce typical equilibrium values for the fluxes : B₀ = V/5 ; C_{Eq} = (8/5) $\tau_C V = 40,000$ PgC. If we assume a relaxation time τ_C of 200 kyr (Archer et al., 1997), we obtain V = (5/8) C_{Eq}/ $\tau_C = 125$ TgC/yr and B₀ = 25 TgC/yr. For a larger value $\tau_C = 400$ kyr (Archer et al. 2005), we would get V = 62 TgC/yr. There is no consensus on the actual total carbon emissions from volcanism (including all aerial and submarine sources), but these values for V (or τ_C) span more or less the range of current

estimates from about 40 to 175 TgC/yr (Burton et al, 2013). It must be stressed that B stands for all organic carbon fluxes, whether they correspond to organic carbon burial (positive contributions to B) or to organic matter oxidation (negative contributions to B). If the long-term average equilibrium value B₀ needs to be positive to account for the isotopic balance as shown above, this is not necessary always the case for the

110 instantaneous values of B, as we will illustrate it below with the astronomical forcing. Indeed, B represents a sum of positive and negative terms whose individual absolute magnitudes are much larger than the long-term net value B₀. For instance, the oxidation of petrogenic organic carbon alone will contribute negatively to B, with a magnitude that may be as large as 40 TgC/yr (Blair et al., 2003).

The isotopic ¹³C budget can be written as:

120

115 $d/dt(C \ \delta^{13}C) = V \ \delta^{13}V - B \ \delta^{13}B - D \ \delta^{13}D$

where $\delta^{13}C$ is the isotopic composition of ocean carbon, $\delta^{13}V$ the isotopic composition of the volcanic carbon input, $\delta^{13}B$ the isotopic composition of organic matter and $\delta^{13}D$ the isotopic composition of marine carbonates. This can be re-written as: $C (d\delta^{13}C/dt) + (dC/dt) \delta^{13}C = V \delta^{13}V - B \delta^{13}B - D \delta^{13}D$

or
$$C (d\delta^{13}C/dt) = V \delta^{13}V - B \delta^{13}B - D \delta^{13}D - (V - B - D)\delta^{13}C$$

= V (
$$\delta^{13}$$
V- δ^{13} C) - B (δ^{13} B- δ^{13} C) - D (δ^{13} D - δ^{13} C)

If we neglect isotopic fractionation during carbonate precipitation (in other words, $\delta^{13}D = \delta^{13}C$) and more generally during carbonate compensation, we finally obtain:

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(2b) $d\delta^{13}C/dt = (V(\delta^{13}V - \delta^{13}C) - B(\delta^{13}B - \delta^{13}C))/C$

Didier Paillard 8/8/y 11:42 Commentaire [2]: Rev.#2, RC1

Didier Paillard 8/8/y 13:26 Commentaire [3]: Rev.#2, RC4

Commentaire [4]: Rev.#1

Didier Paillard 8/8/y 11:54 Commentaire [5]: Rev.#2, RC3 In the following we will assume a constant -5% volcanic source $\delta^{13}V$, as well as a constant -25‰ organic matter value $\delta^{13}B$

125 (eg. Porcelli and Turekian, 2010).

In order to translate the total carbon content C into an equivalent pCO₂ level, we will use a simple scaling. Indeed, if we assume, to first order, that C may represent the carbon content of a well-mixed ocean, then from chemical equilibrium pCO₂ should be proportional to $[HCO_3^-]^2/[CO_3^{2-}]$. After carbonate compensation (i.e. assuming that $[CO_3^{2-}]$ remains constant) and considering that C is dominated by bicarbonates $[HCO_3^-]$ under standard pH conditions, we end up with the approximate

- 130 scaling that pCO₂ varies roughly as C², or pCO₂ = 280 (C/40,000)² (in ppm). To reproduce a multi-million year trend, we need to add one explicitly in the weathering relaxation: $W = C/\tau_C = (\Delta C + C_{Eq} - \gamma t)/\tau_C$, with the coefficient γ set to 1,2 TgC/yr to obtain the desired pCO₂ levels at the start of the simulation, ie. about 350 ppm at 4 MyrBP, according to current estimates (Bartoli et al., 2011; Seki et al., 2010). The model is integrated from an arbitrary initial condition at 5 MyrBP and the first 1 Myr is discarded.
- 135 In the following, we are describing how carbon burial B should vary with monsoons, and what consequences these variations have on the total carbon content C as well as on carbonate isotopes δ^{13} C. In order to represent the monsoon's response to astronomical forcing, we introduce a simple truncation of the precessional forcing:

 $F_0(t) = \max(0, -e\sin\omega)$

where e is the eccentricity and ω the climatic precession.

140 Indeed, soil erosion or sediment transport are dominated by intense events, not by the average climate. Such a non-linear response can be mimicked in a simple way by the above expression that accounts only for positive monsoonal forcing, not for negative one. Consequently, the model will be influenced by the amplitude modulation of the precessional forcing, ie. the eccentricity. To avoid useless parameters, we furthermore introduce the normalization:

 $F = F_0 / Max(F_0) - \langle F_0 / Max(F_0) \rangle$

 $145 \quad \mbox{which results in a precessional forcing } F(t) \mbox{ with amplitude one and zero mean.}$

We implicitly account for a slow terrestrial organic carbon reservoir (soil) as "buried organic carbon". It is reasonable to assume that monsoon, or enhanced precipitation will favor primary production and soil formation. But this recent soil together with older soils and with petrogenic organic carbon (Galy et al., 2008) will be eroded and transported to the ocean through enhanced river discharges. If the corresponding carbon is remineralized in the ocean without too much burial in the

150 alluvial fan, the net perturbation of the burial flux is likely to be negative (ie. net "old" soil erosion and remineralization). We will refer to this case as the "Amazon-like" situation, with the perturbation F(t) being subtracted to the baseline burial B_0 by writing $B = B_0 - a F(t)$. In contrast, if most of the organic carbon is buried and preserved in the sediment, then the perturbation is likely to be positive, since it induces a net "recent" soil formation and burial. We call this the "Himalayan-like" situation, with now $B = B_0 + a F(t)$. Before 1 MyrBP and the associated major sea level changes, the river fans and

155 continental shelves should evolve mostly in a progradational way (see scheme on Fig. 3), a situation which a priori favors

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Didier Paillard 8/8/y 13:29 Commentaire [6]: Instead of "fractionation". Cf. Peter Köhler's comment #1

Didier Paillard 8/8/y 11:58 Commentaire [7]: Rev.#2, RC2

& Peter Köhler's comment #1 Didier Paillard 8/8/y 11:59

Commentaire [8]: Peter Köhler's comment #3

Didier Paillard 8/8/y 11:58 Commentaire [9]: Rev.#2, RC4 & Peter Köhler's comment #2

Didier Paillard 9/8/y 17:19 Commentaire [10]: Peter Köhler's comment #5

Didier Paillard 9/8/y 17:22 Commentaire [11]: Peter Köhler's comment #4

Didier Paillard 8/8/y 12:08 Commentaire [12]: Rev.#1 organic carbon remineralization, while aggradational situations are likely to be more frequent in the late Pleistocene, with therefore a possible temporary reversal of the organic carbon burial.

3 Results

Our first simulations, with $B = B_0 - a F(t)$, correspond to a perpetual "Amazon-like" situation. They correspond to

- 160 experiment *a* (black lines) with no trend in the total carbon, and experiment *b* (blue lines), with an explicit linear trend in carbon. [The value of the parameter *a* is chosen in order to obtain approximately the correct amplitude for these simulated 400-kyr oscillations (a = 50 TgC/yr). Still, as can be seen on Fig.2, we obtain a surprisingly good match between the simulated and observed δ^{13} C, with overall very similar cycles. More specifically, the δ^{13} C black and blue simulated curves are superimposed and almost undistinguishable, since the linear trend added to the carbon cycle has almost no impact on the
- 165 δ^{13} C. They are both most of the time within the range of observed values (gray curves). The two main exceptions occur at about 0.3 and 2.3 MyrBP, with the simulated δ^{13} C being significantly too high. In experiment *a* (black lines), pCO₂ is oscillating around its equilibrium value, with two significant negative excursions occurring near 2.5 MyrBP and near 0.5 MyrBP. These lower values are directly linked to the ~2.4 Myr modulation of eccentricity. Obviously, with fixed or periodic parameters, this model cannot simulate a long term decreasing trend in carbon. When explicitly adding such a linear
- 170 decreasing trend (experiment b, blue lines), the two minima described above become two decreasing steps. The first one, occurring around 2.8 MyrBP, is coincident with the Plio-Pleistocene transition and the development of Northern hemisphere glaciations. The second one near 0.8 MyrBP is coincident with the Mid-Pleistocene transition (MPT) and the significant amplification of glaciations. Note that the timing of these two steps is directly linked to the astronomical forcing: it does not depend at all on the specifics of the trend that we used here. Two similar pCO₂ decreasing episodes are also seen in the data
- 175 (Figure 1) though it is difficult to associate them with a precise timing, due to the difficulties to reconstruct accurately pCO_2 from indirect proxies.

In order to account for the observed departure of the δ^{13} C oscillations from a simple eccentricity forcing, we need to introduce a retroaction of Quaternary sea level changes onto the sedimentary dynamics of alluvial fans and continental shelves, and consequently onto organic carbon burial. As explained above, we will reverse the sign of our burial flux perturbation, and change it into $B = B_0 + a F(t)$ when some conditions are met on the geomorphology of river outputs. In

- 180 perturbation, and change it into $B = B_0 + a$ F(t) when some conditions are met on the geomorphology of river outputs. In particular it is necessary to account for a changing reservoir size that can be filled with sediments in an aggradational way. Indeed, at the first major sea level drop, rivers are incising though the river and fan bedrock, thus providing room for the accumulation of sediments loaded with organic carbon. This volume should be filled progressively with sedimentary organic carbon up to a point when further river incision, and consequent aggradation of sediment, do not affect the global organic
- 185 carbon anymore but only move sedimentary carbon from one place to the other. In other words, we will assume that the global "Himalayan-like" situation (ie. net organic carbon burial) is only a transient situation, linked to the first occurrence of a sea level minima. In order to illustrate this mechanism, we add a new equation for the slow geomorphological reservoir S

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Didier Paillard 8/8/y 12:13 Commentaire [13]: Rev.#2, RC5

Didier Paillard 8/8/y 12:20 Commentaire [14]: Rev.#2, RC13 for organic carbon in river beds or river fans. We define its maximal size S_{MAX} from the observed sea level changes using the reference stack LR04 (Lisiecki and Raymo, 2005) by finding the previous sea level minima z_{MIN} (ie. the lower envelope) with the scaling $S_{MAX} \sim z_{MIN}^3$ since it represents a volume of sediment (see Fig. 2).

In other words, the sedimentary organic carbon reservoir S grows at the pace of the above mentionned astronomical perturbation F₀(t) up to its maximal size S_{MAX}. After a short transient period, this reservoir remains therefore equal to this maximum value S_{MAX} in the absence of major sea level drops, as during the pre-Quaternary period. In contrast, for each significant sea level drop, S_{MAX} increases abruptly and we start a new transient phase whose duration is linked to parameter *b*. When S is small compared to the maximal reservoir size S_{MAX}, then the aggradiational scheme is favoured, with river beds and deltaïc net organic carbon burial. But when S is close to its maximum value, we switch back to a mostly progradational sedimentation scheme, meaning that potential sea level changes will no more affect net global organic carbon burial :

200

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(3b)

 $\mathbf{B} = \mathbf{B}_0 - a \mathbf{F}(\mathbf{t})$

 $B = B_0 + a F(t)$

Using this simple crude criteria, we obtain the results show on Fig. 2 (experiment *c*, red lines). As expected, this simple model does switch from the background "Amazon-like" or progradational burial mode to a "Himalayan-like" or aggradational mode, after each significant sea level drop, and most notably at two time periods, the first one between 2.4 and

- 205 2.5 MyrBP (as a consequence of the Plio-Pleistocene transition) and the second and largest one between 350 and 650 kyrBP (as a consequence of the MPT). The start of these transient periods is directly linked to sea level drops, according to the LR04 forcing, while the duration of these transients is linked both to the 0.85 S_{MAX} threshold and the *b* parameter, whose values are chosen to qualitatively better match the δ^{13} C data. For the results show on Fig.2, *b* = (160 kyr)⁴. Indeed, this sedimentary switch mechanism allows for a much better agreement with measured δ^{13} C around 0.3 and 2.3 MyrBP, while
- 210 the first simulations were systematically too high at this time, as illustrated by the difference between the blue and red curves on Fig.2. We also simulate correctly the δ^{13} C maximum around 500 kyrBP and the occurence of two broad "500 kyr" cycles over the last million years. With this burial mode switching mechanism, we are also able to predict an absolute minimum in carbon content, or long-term pCO₂, around 600 kyrBP, in rather good agreement with the long term trend of pCO₂ measured in Antarctic ice cores. Indeed, pCO₂ from the Dome C record is about 5 to 10 ppm lower before the MPT (between 400 and
- 215 800 kyrBP), which is also what we obtain in our experiment *c*.

if $S < 0.85 S_{MAX}$:

otherwise:

3 Discussion

When variations in B, as determined by parameter *a*, are smaller than the baseline value B₀, the model cannot reproduce the oceanic amplitude of δ^{13} C observed in marine benthic records. The observed 400-kyr signal in δ^{13} C records therefore requires major changes in the organic carbon burial, with almost no global net burial, but net oxidation episodes, during

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Didier Paillard 8/8/y 13:34 Commentaire [15]: Rev.#2, RC6

Didier Paillard 8/8/y 15:42 **Commentaire [16]:** Rev.#2, RC6 Didier Paillard 9/8/y 13:15 **Commentaire [17]:** Rev.#2, RC7 Didier Paillard 8/8/y 15:52

Commentaire [18]: Rev.#2, RC8 Didier Paillard 9/8/y 13:18

Supprimé: In order to further improve the

qualitative match between measured and simulated $\delta^{13}C$, we can also add artificially a component linked to glacial-interglacial cycles. This is done by adding the detrended sea level LR04 curve to $\delta^{13}C$ obtained from experiment c (orange line on figure 2). This allows to account for the significant 100-kyr periodicity seen in the carbon isotopic record, and usually attributed to glacial-interglacial changes in the global biospheric size (eg. Shackleton, 1977). - Didier Paillard 8/8/y 15:40

Supprimé: In order to reproduce the observed amplitude of the 400-kyr oscillations in the δ^{13} C records (about 0.5%), the strength of the forcing *a* needs to be of the same order than the baseline value B_0 This is a robust feature, which does not depend on model setting or parameters -

Didier Paillard 8/8/y 12:15 Commentaire [19]: Rev.#2, RC5 maxima of precessional forcing. This strong forcing therefore implies significant oscillations in the Earth carbon cycle for this time frequency, up to 4 or 5% in total carbon content. This is translated here into 10 to 20 ppm variations of pCO_2 using our simple scaling, but it is very likely that these changes would be much larger, when accounting for interactions between pCO_2 and climate. Indeed, colder climates are more favorable to oceanic carbon storage, as observed during the last glacial

240 cycles. According to this mechanism, in the ordinary sedimentary regime (progradation), we obtain changes in the carbon cycle with pCO₂ maxima and δ^{13} C minima associated directly to eccentricity maxima. This is indeed consistent with long Cenozoïc records (eg. Pälike et al, 2006).

When we allow for changes in the sedimentary regime triggered by sea level changes, the model can also reproduce more peculiar features. Indeed, up to now it has been difficult to explain the last two long-term cycles observed in the marine δ^{13} C,

245 each being approximately 500 kyr-long, with a maximum now (δ¹³Cmax-I), a well-marked maximum at about 500 kyr BP (δ¹³Cmax-II) and a previous one around 1000 or 1100 kyr BP (δ¹³Cmax-III). In the model described here, these two long oscillations are generated from the eccentricity forcing, but with an abrupt switch to aggradation mode at about 620 kyrBP caused by the sea level drop at MIS 16. This switch reverses the phase of the 400-kyr carbon oscillation during a few hundred thousands of years. Interestingly, this also induces a slight minimum in the carbon (or pCO₂) results, consistent with the observed low pCO₂ values observed in the Antarctic ice core around 600-700 kyrBP. [This mechanism also allows for simulated marine δ¹³C in better agreement with data at about 2.4 MyrBP.]

It was already noted (Wang et al., 2004) that the climatic evolution since the last million years, in particular the Mid-Pleistocene Transition (MPT, about 0.8 MyrBP) and the Mid-Brunhes Event (MBE, about 0.4 MyrBP) were associated with the carbon isotopic maxima δ^{13} Cmax-II and δ^{13} Cmax-III. This was a strong indication of a possible causal link between the

- 255 long-term well-recognized eccentricity forcing on the carbon cycle and the Plio-Pleistocene climatic evolution. There is therefore a strong incentive to build a mechanistic astronomical theory of the carbon cycle. But a prerequisite towards understanding this long-term precessionally forced carbon cycle and its climatic consequences is to explain the observed changes during the Quaternary, in terms of δ^{13} C and simultaneously in the atmospheric CO₂ levels (Lüthi et al, 2008). The model results outlined above are a first step in this direction.
- 260 As detailed above, the fact that the 400-kyr carbon isotope cycle is perturbed during the Pleistocene strongly points towards a major role for organic matter burial over continental shelf areas being affected by sea-level changes. Obviously, this model is far too simple to represent faithfully the complexities of sedimentary dynamics in coastal areas, its consequences on organic matter preservation, on carbon cycle and ultimately on climate. Besides, we provide here no explanation for the prescribed multi-million year decreasing trend in pCO₂. There is unfortunately no clear consensus on the actual mechanisms
- 265 involved, though this trend has been often attributed to long-term changes in continental weathering linked either to mountain uplift (Raymo and Ruddiman, 1992), to continental drift or mantle degassing rate (Lefebvre et al., 2013). Furthermore, we considered only sea-level changes as a potential feedback on organic matter burial in coastal areas. Obviously, many other important climatic feedbacks would also play a role: for instance increased temperatures would

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Didier Paillard 9/8/y 14:32 Commentaire [20]: Rev.#2, RC10 Didier Paillard 9/8/y 13:28 Commentaire [21]: Rev.#2, RC8 probably reduce net primary production as a consequence of increased stratification, and therefore reduce organic carbon

270 deposition in coastal sediments, but it would also decrease oxygen concentrations and consequently would favor organic matter preservation. Similarly, stronger monsoons events would enhance the delivery of nutrients to the continental shelves, and therefore biological productivity. This would in addition deliver more fine-grained clay minerals that are necessary to seal and preserve organic matter from oxidation. This would work opposite to our continental soil-carbon mechanism for which enhanced monsoons lead to more organic carbon oxidation in agreement with the isotopic records. But, as a proof of concept, our model is chosen as minimalistic as possible. It does not attempt to include all potentially important mechanisms.

275 concept, our model is chosen as minimalistic as possible. It does not attempt to include an potentially important mechan

4 Conclusion

Our basic assumptions are primarily based on recent re-assessments of riverine organic carbon inputs to the ocean. With the above conceptual model, we demonstrate that simple mechanistic assumptions can account for the major patterns of the observed global evolution of carbon and carbon isotopes over this time period: First, enhanced precessional forcing linked to

- 280 high eccentricity leads to more continental organic carbon been washed out and remineralized, therefore a net decrease in overall organic carbon burial. Second, this mechanism is temporarily reversed following major sea-level drops associated with glaciations. [This model was built on the premises that changes in organic matter or petrogenic organic carbon fluxes are responsible for the 400-kyr oscillations observed in Cenozoïc ¹³C records, and that the large sea-level variations occurring during the Quaternary are strongly affecting this process. Continental margins and sedimentary fans are a very likely key
- 285 component, as illustrated by our simple conceptual model. But obviously, many complex processes are involved in the interactions between organic matter burial or oxidation, monsoons and sea-level changes. The geomorphological mechanism described here is one possibility which allows, for the first time, to account both for the persistent 400-kyr oscillation observed in ¹³C records during the Cenozoïc, but also for its change during the last million years. It also suggests the occurrence of possibly significant CO₂ drops at about 0.8 MyrBP (Mid-Pleistocene transition) and at about 2.8 MyrBP (Plio-
- 290 Pleistocene transition), that would ultimately link the timing of these transitions to the astronomical forcing. Our model also provides a possible explanation for the puzzling shifted level in the CO₂ records associated with the MBE

References

Archer, D., Kheshgi, H., Maier-Raimer, E., Multiple timescales for neutralization of fossil fuel CO2. Geophys. Res. Lett (1997) vol. 24 (4) pp. 405-408.

295 Archer, D., Fate of fossil fuel CO₂ in geologic time. J.G.R. (2005) vol. 110 (C09S05) pp. doi:10.1029/2004JC002625 Bartoli, G., Hönisch, B., Zeebe R.E., Atmospheric CO₂ decline during the Pliocene intensification of Northern Hemisphere glaciations. Paleoceanography (2011) vol. 26 (4)

9

Didier Paillard 9/8/y 14:15 Commentaire [22]: Rev.#1

Didier Paillard 9/8/y 14:29 Commentaire [23]: Rev.#2, RC9 Bassinot, F., Beaufort, L., Vincent, E., Labeyrie, L., Rostek, F., Müller, P., Quidelleur, X., Lancelot, Y., Coarse fraction fluctuations in pelagic carbonate sediments from the tropical Indian Ocean: A 1500-kyr record of carbonate dissolution. Paleoceanography (1994) vol. 9 (4) pp. 579-600

Billups, K., H. Pälike, J. Channell, J. Zachos, and N. Shackleton (2004), Astronomic calibration of the late Oligocene through early Miocene geomagnetic polarity time scale, Earth Planet. Sci. Lett., 224, 33-44.

300

Blair, NE, Leithold, EL, Ford, ST, Peeler, KA, Holmes, JC, Perkey, DW, The persistence of memory: the fate of ancient sedimentary organic carbon in a modern sedimentary system. Geochimica et Cosmochimica Acta (2003) vol. 67 pp. 63

305 Burdige, D. J. Burial of terrestrial organic matter in marine sediments: A re-assessment. Glob. Biogeochem. Cycles 19, GB4011 1–7 (2005).

Burton MR, Sawyer GM, Granieri D (2013) Deep carbon emissions from volcanoes. Reviews in Mineralogy and Geochemistry 75: 323-354

Cramer, B., Wright, J.D., Kent, D.V, Aubry, M.-A., Orbital climate forcing of δ13C excursions in the late Paleocene–early
Biocene (chrons C24n–C25n). Paleoceanography (2003) vol. 18 (4) pp. 25

Galy, V., France-Lanord, C., Beyssac, O., Faure, P., Kudrass, H., Palhol, F., Efficient organic carbon burial in the Bengal fan sustained by the Himalayan erosional system. Nature (2007) vol. 450 (7168) pp. 407-410

Galy, V., Beyssac, O., France-Lanord, C., Eglinton, T. Recycling of Graphite During Himalayan Erosion: A Geological Stabilization of Carbon in the Crust. Science (2008) vol. 322 (5903) pp. 943-945

315 Hilgen F.J. et al. Present status of the astronomical (polarity) time-scale for the Mediterranean Late Neogene. Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences (1999) vol. 357 (1757) pp. 1931-1947

Hilton, R.G., Galy, V., Gaillardet, J., Dellinger, M., Bryant, C., O'Regan, M., Gröke, D.R., Coxall, H., Bouchez, J., Calmels,
 D., Erosion of organic carbon in the Arctic as a geological carbon dioxide sink. Nature (2015) vol. 524 (7563) pp. 84-87.

320 Hönisch, B., Hemming, N.G., Archer, D., Siddall, M., McManus, J., Atmospheric Carbon Dioxide Concentration Across the Mid-Pleistocene Transition. Science (2009) vol. 324 (5934) pp. 1551-1554

Kuiper, K.F., Deino, A., Hilgen, F.J., Krijgsman, W., Renne, P.R., Wijbrans, J.R., Synchronizing Rock Clocks of Earth History. Science (2008) vol. 320 (5875) pp. 500-504.

Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A.C.M., A long-term numerical solution for the insolation quantities of the Earth. Astronomy and Astrophysics (2004) vol. 428 pp. 261-285, DOI: 10.1051/0004-6361:20041335

Lefebvre, V., Donnadieu, Y., Goddéris, Y., Fluteau, F., Hubert-Théou, L., Was the Antarctic glaciation delayed by a high degassing rate during the early Cenozoic?. Earth Planet. Sci. Lett. (2013) vol. 371-372 (C) pp. 203-211

Lisiecki, L., Raymo, M., A Pliocene-Pleistocene stack of 57 globally distributed benthic d¹⁸O records. Paleoceanography (2005) vol. 20 pp. PA1003, doi:10.1029/2004PA001071

330 Lourens, L.J., Antonarakou, A., Hilgen, F.J., VanHoof, A.A.M., Vergnaud-Grazzini, C., Zachariasse, W.J., Evaluation of the Plio-Pleistocene astronomical timescale. Paleoceanography (1996) vol. 11 (4) pp. 391-413 Lüthi, D., Le Floch, M., Bereiter, B., Blunier, T., Barnola, J.-M., Siegenthaler, U., Raynaud, D., Jouzel, J., Fisher, H., Kawamura, K., Stocker, T., High-resolution carbon dioxide concentration record 650,000–800,000 years before present. Nature (2008) vol. 453 (7193) pp. 379-382.

Mix, A. C., N. G. Pisias, W. Rugh, J. Wilson, A. Morey, and T. K. Hagelberg (1995), Benthic foraminifer stable isotope record from Site 849 (0 – 5 Ma): Local and global climate changes, Proc. Ocean Drill. Program Sci. Results, 138, 371 – 412. Paillard, D., Donnadieu, Y., A 100 Myr history of the carbon cycle based on the 400 kyr cycle in marine δ¹³C benthic records. Paleoceanography (2014)

Paillard, D., Quaternary glaciations: from observations to theories. Quat. Sci. Rev. (2015) vol. 107 (C) pp. 11-24

Pälike, H., R. D. Norris, J. O. Herrle, P. A. Wilson, H. K. Coxall, C. H. Lear, N. J. Shackleton, A. K. Tripati, and B. S. Wade (2006), The heartbeat of the Oligocene climate system, Science, 314, 1894–1898.
 Porcelli, D. and Turekian, K.K., The History of Planetary Degassing as Recorded by Noble Gases, §6.6.1 in Readings from the Treatise on Geochemistry, edited by Holland, H.D. and Turekian, K.K., (2010)
 Bauma, M. Buddiman, W.F. Tastonia forzing of late comparis climate. Nature (1002) vol. 250 np. 117, 122

Raymo, M., Ruddiman, W.F., Tectonic forcing of late cenozoic climate. Nature (1992) vol. 359 pp. 117-122.

345 Rickaby, R., E. Bard, E. Sonzogni, F. Rostek, L. Beaufort, S. Barker, G. Rees, and D. Schrag (2007), Coccolith chemistry reveals secular variations in the global ocean carbon cycle?, Earth Planet. Sci. Lett., 253, 83–95. Russon, T., D. Paillard, and M. Elliot (2010), Potential origins of 400–500 kyr periodicities in the ocean carbon cycle: A box

model approach, Global Biogeochem. Cycles, 24, GB2013, doi:10.1029/2009GB003586.

Seki, O., Foster, G.L., Schmidt, D.N., Mackensen, A., Kawamura , K., Pancost, R.D., Alkenone and boron-based Pliocene
 pCO₂ records. Earth Planet. Sci. Lett. (2010) vol. 292 (1-2) pp. 201-211

Sexton, P., R. D. Norris, P. A. Wilson, H. Pälike, T. Westerhold, U. Röhl, C. T. Bolton, and S. Gibbs (2011), Eocene global warming events driven by ventilation of oceanic dissolved organic carbon, Nature, 471, 349–352.

Shackleton, N. J., M. A. Hall, and D. Pate (1995), Pliocene stable isotope stratigraphy of Site 846, Proc. Ocean Drill. Program Sci. Results, 138, 337–355.

355 Wang, P., J. Tian, and L. Lourens (2010), Obscuring of long eccentricity cyclicity in Pleistocene oceanic carbon isotope records, Earth Planet. Sci. Lett., 290, 319–330.

Wang, P., J. Tian, X. Cheng, C. Liu, and J. Xu (2004), Major Pleistocene stages in a carbon perspective: The South China Sea record and its global comparison, Paleoceanography, 19, PA4005, doi:10.1029/2003PA000991.

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Figure 1: From top to bottom: — pCO₂ records from Antarctic ice cores (purple: Lüthi et al, 2008); from boron isotopes in marine cores (orange: Hönisch et al, 2009; light blue: Bartoli et al., 2011) and alkenone isotopes (pink and blue lines for the min and max envelope, from Seki et al, 2010). — 8¹³C in cores 1143 (red: Wang et al., 2004); 849 (blue: Mix et al., 1995); 846 (green: Shackleton et al., 1995). — the same 8¹³C records filtered at 400-ky (bandpass = 2.5 Myr⁻¹) — eccentricity (grey, from Laskar et al., 2004) and filtered eccentricity (black).





Figure 2: From top to bottom: — Precessional forcing $F_0(t) = Max(0, -e \sin\omega)$ (black line) from Laskar et al. (2004). — Sea level curve LR04 (black line) from Lisiecki and Raymo (2005) used to compute the river incision z_{MIN} defined as the previous sea level minima (blue line). — The geomorphological variable *s* used from experiment *c* (red lines) relaxed to its prescribed maximum value $s_{MAX} \sim z_{MIN}^3$ (black line). The orange shaded areas correspond to the aggradation regimes (ie. s < 0.85 s_{MAX}). — Total carbon C rescaled as pCO₂ for experiments *a* (black, precessional forcing only), *b* (blue, idem, with a linear trend in carbon), and *c* (red, using the geomorphological dynamics from equation (3)) — Carbon isotopic composition δ^{13} C for experiments *a* (black), *b* (blue), and *c* (red). In grey, the min and max values of the ¹³C records from Figure 1. — The 400-kyr filtered values of δ^{13} C results (blue

and te (red), an grey, ite lim and max values of the corrections from Figure 1. In order to hold in the same of filtered records (grey). — The 400-kyr filtered eccentricity as in Fig.1. In order to obtain these results, we choose $\pi_c = 200$ kyr (Archer et al., 1997) or equivalently V = 125 TgC/yr. The trend (experiments *b* and *c*) is set to y = 1,2 TgC/yr to induce a drift from about 350 ppm to about 280 ppm. The amplitude of the organic matter burial perturbation (experiments *a*, *b* and *c*) is set to *a* = 50 TgC/yr. The filling rate of the sedimentary reservoir (experiment *c*) is set to *b* = (160 kyr)⁻¹. The model is integrated from an arbitrary initial condition at 5 MyrBP and the first 1 Myr is discarded.

Didier Paillard 9/8/y 16:56 **Supprimé:** In orange, experiment *c* with the addition of glacial-interglacial variability scaled on sea level. Didier Paillard 9/8/y 16:57

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Figure 3: Simple scheme of the two different geomorphological dynamics considered here : a/ With small sea level changes, we assume that the dominant sedimentary regime is progradation, with rather small organic carbon burial in coastal areas. The net effect of precessional forcing is (old) soil erosion, therefore a net transfer of carbon to the ocean-atmosphere ; b/ With large sea level changes during the late Quaternary, the dominant sedimentary regime can switch temporarily to aggradation just after major sea level drops and river incisions. During these transitory phases, the net effect of precessional forcing is reversed, with net organic carbon burial in river beds and river fans.