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The role of palaeogeography in the Phanerozoic history of atmospheric CO_2 and climate

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ABSTRACT

The role of the palaeogeography on the geological evolution of the global carbon cycle has been suspected since the development of the first global geochemical models in the early 80s. The palaeogeography has been rapidly recognized as a key factor controlling the long-term evolution of the atmospheric CO_2 through its capability of modulating the efficiency of the silicate weathering. First the role of the latitudinal position of the continents has been emphasized: an averaged low latitudinal position promotes the CO_2 consumption by silicate weathering, and is theoretically associated to low CO_2 periods. With the increase of model complexity and the explicit consideration of the hydrological cycle, the importance of the continentality factor has been recognized: periods of supercontinent assembly coincide with high pCO_2 values due to the development of arid conditions which weaken the silicate weathering efficiency. These fundamental feedbacks between climate, carbon cycle and tectonic have been discovered by pioneer modelling studies and opened new views in the understanding of the history of Earth's climate. Today, some of the key features of the Phanerozoic climate can be explained by: (1) continental drift; (2) small continental blocks moving to tropical belts; and (3) modulation of the climate sensitivity to CO_2 by palaeogeography changes. Those results emphasize the need for a careful process-based modelling of the water cycle and climate response to the continental drift.

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1. Introduction

Palaeogeography and Earth climate are tightly linked. As noticed by Hay et al. (1990), the existence of a link between the latitudinal

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distribution of land mass and the continental climate has been first postulated by Lyell (1830), almost a century before the publication of the continental drift theory (Wegener, 1912). Assuming a constant land surface, Lyell (1830) proposes that filling the Arctic region with emerged continents at the expense of equatorial territories would result in a severe polar glaciation and in a drop of the global temperature. Then comes the continental drift theory (Wegener, 1912) and the prediction of Köppen and Wegener (1924) that if all continents are located in the equatorial area for a given past period, continental and shelfal markers of climate will all indicate warm conditions, making a strong case for a link between palaeogeography and climate.

Since the end of the seventies, an abundant literature deals with the link between continental configuration and climate. Land mass distribution modifies the regional and global climate (Gyllenhaal et al., 1991), through changes in the oceanic circulation (Kennett, 1977; von der Heydt and Dijkstra, 2006; Lagabrielle et al., 2009 and reference therein; Lefebvre et al., 2012; Zhang et al., 2012), albedo feedbacks (Hay et al., 1990; Horton et al., 2010), and direct impact on the precipitation/evaporation balance (Barron et al., 1989; Otto-Bliesner, 1995). In addition to geological features, climate modelling studies explore the role of the continental configuration on pCO_2 thresholds needed to explain ice/green-house successions (Barron and Washington, 1982; Crowley and Baum, 1991, 1995; Barron et al., 1993; Gibbs et al., 1997; Hyde et al., 1999; Herrmann et al., 2004; Donnadieu et al., 2006; Craggs et al., 2012; Horton et al., 2012; Spicer et al., 2008). However all these studies share a common characteristic: they prescribe atmospheric greenhouse gas levels. Specifically, CO_2 level is a boundary condition and is thus assumed to be an external forcing of the climate system, whatever the continental configuration.

But things are more complex. Since weathering of Ca–Mg silicates of crustal rocks consumes atmospheric CO_2 at a rate depending on the Earth's climate, CO_2 levels are necessarily depending on the palaeogeography at long-term scale. Basically, the rock mineral dissolution directly depends: (1) on the availability of water, thus on the continental area located within humid climatic belts; and (2) on the temperature, thus on the latitudinal distribution of the continents. Surprisingly, the interplay between palaeogeography and global carbon cycle has not been often studied in the recent years, despite the fact that it may be potentially a first order forcing of climate at the geological timescale. Instead of working on the geography effect, most of the efforts have been concentrated on the role of land plants (Berner, 2004), and of mountain building and physical erosion (Raymo et al., 1988; Raymo, 1991).

In this contribution, we propose an overview of the studies devoted to the impact of palaeogeography on the long-term carbon cycle ($>10^5$ yrs). We show that continental precipitations and runoff are highly sensitive to the palaeogeography. We demonstrate that the continental configuration is a first order factor controlling the atmospheric CO_2 level and the Earth's climate evolution.

2. Pioneer modelling studies of the impact of palaeogeography on the global carbon cycle

According to the Earth thermostat theory (Walker et al., 1981), for a given solid Earth CO_2 degassing, the total CO_2 consumption by continental weathering should be the same, whatever the location of continents. This is a direct consequence of the short response time of carbon and alkalinity in the ocean–atmosphere system, respectively 200 and 3 kyr (François and Goddéris, 1998) and of the negative feedback exerted by climate on weathering (Walker et al., 1981). The CO_2 consumption by silicate weathering is always tracking the CO_2 supply by solid Earth degassing, to avoid too large CO_2 fluctuations and repeated climatic catastrophe (Berner and Caldeira, 1997).

Marshall et al. (1988) first proposed that the palaeogeography might be a key parameter of the evolution of the geological carbon cycle (Table 1). They found that atmospheric CO_2 must be at high levels when most of the continents are located around the poles. Conversely, CO_2 must be low and global climate cold when continents are located around the equator. They argue that this is a direct consequence of a promoted CO_2 consumption by continental silicate rock dissolution when continents are located in the warm equatorial belt. Because temperatures are higher at the equator than at the pole, Marshall et al. (1988) note that the weatherability (i.e., the susceptibility to weathering) is higher when continents are close to the equator. As a consequence, the same total weathering flux will be reached at a much lower CO_2 level compared to a world where continents are located at the pole. Although their study was conceptual with generic continental configurations, Marshall et al. (1988) had already proposed that the late Precambrian glaciations might have been promoted by an equatorial location of the continents.

Worsley and Kidder (1991) further explore the Marshall et al. (1988) hypothesis through additional qualitative considerations. Similarly to Marshall et al., they suggest that a ring world (all continents located along the equator) should be characterized by lower atmospheric CO_2 (and hence colder global conditions) than a cap world (all continents around the poles). Provocatively, Worsley and Kidder (1991) even propose that the averaged latitude of the continents alone specify CO_2 levels.

Marshall and collaborators were the first to couple a process-based climate model with a mathematical description of the continental weathering. The climate model used was quite simple: an energy-balanced model (EBM). This model calculates the Earth surface temperature as a function of latitude accounting for the vertical budget of energy and for the meridional heat transport. Similar models, coupling a description of the carbon cycle with a zonal energy-balanced model, have been used to explore the Phanerozoic CO_2 history (François and Walker, 1992; Goddéris and François, 1995; Goddéris and Joachimski, 2004; Tajika, 2007). However, it is important to note that if EBM models are reliable in terms of temperature predictions, they lack a physical description of the water cycle and, as such, are not able to calculate accurately the runoff, a key parameter of continental weathering (Dessert et al., 2003; Oliva et al., 2003). To overcome this limitation, a new generation of models has been developed.

3. Exploring the role of palaeogeography on the water and carbon cycle

Continental silicate weathering and associated CO_2 sink are heavily dependent on continental runoff, which is the difference between rainfall and evapotranspiration. As such, the relationship between palaeogeographical setting and hydrological cycle must be understood, and this was beyond the abilities of the energy-balanced models previously used. The first study exploring the role of the continent distribution on the hydrologic cycle with an appropriate model was performed in 1989 by Barron and co-authors. They note that the water cycle is an essential part of the geochemical cycles (weathering, erosion, sedimentation), and that too much attention has been paid to temperature reconstructions at the expense of water cycle reconstructions. They explore the role of palaeogeography on the water cycle with the NCAR Community Climate Model (spatial resolution of 7.5° long × 4.5° lat). Contrary to energy-balanced models, the CCM includes a calculation of evaporation and condensation, and cloud cover. Such 3D climate models (also called General Circulation Models or GCM) explicitly simulate the water cycle, calculating the evaporation and rainfall on a 2-D map of the world, and the three-dimensional transport of water vapour in the atmosphere. Their main finding is that continental positions can strongly alter the hydrologic cycle (rainfall and runoff) if their positions affect the supply of moisture to the atmosphere. This is particularly the case in a simulation assuming a continuous belt of continents between 15°S and 15°N. Compared to the present day, this ring world is 8 °C globally warmer, but global rainfall is maintained at its present day rate because the area of intense evaporation along the equator is lacking oceanic surfaces. Because of this intense continental evaporation, continental runoff

Table 1
A synthesis of numerical models exploring the role of palaeogeography on the geological carbon cycle.

Model	Palaeogeography	Mechanistic calculation of the air temperature	Mechanistic calculation of the water cycle component	Spatial resolution	Parametric calculation of the air temperature	Parametric calculation of the continental runoff
Marshall et al. (1988)	Conceptual	X	X	Zonal		
Barron et al. (1989)	Conceptual and realistic	X	X	7.5°long × 4.5° lat (atmosphere) slab ocean		
Worsley and Kidder (1991)	Conceptual					
Otto-Bliesner (1995)	Realistic	X	X	11.6°lat × 11.25°long (atmosphere) slab ocean		
Gibbs et al. (1997)	Realistic	X	X	4.5°lat × 7.5°long (atmosphere) slab ocean		
Berner (2004) and all GEOCARB versions				Global, no spatial resolution	X	X
Donnadieu et al. (2004)	Realistic	X	X	10°lat × 50° long (atmosphere) zonal dynamic ocean		
Donnadieu et al. (2006, 2009), Goddéris et al. (2008a, 2008b), Nardin et al. (2011) GEOCLIM model	Realistic	X	X	4.5°lat × 7.5°long (atmosphere) slab ocean		

decreases by a factor of 2 in this extreme ring configuration, with respect to present day runoff. Despite lower continental rainfall, a world with all the continents located beyond 50° latitude (polar location, cap world) displays a much higher runoff rate. These results contradict the finding of Worsley and Kidder (1991). If weathering is firstly dependent on runoff, then a cap world would be more weatherable than a ring world, thus characterized by a lower CO_2 level.

Otto-Bliesner (1995) further investigates the palaeogeographical forcing of runoff by running the NCAR 3D atmospheric general circulation model (i.e., AGCM) of the climate for 14 time slices spanning the Phanerozoic. The palaeogeographical reconstructions are taken from Scotese (www.scotese.com). Aiming at isolating the palaeogeographical effect on the global climate with a focus on the water cycle, Otto-Bliesner (1995) has fixed the CO_2 level at 280 ppmv, the solar constant at its present day value, and has assumed flat continents (no mountains) covered by rocky deserts. She finds that the relationship between runoff and palaeogeography is critical but more complex than the Marshall et al. (1988) and Worsley and Kidder (1991) simple scheme. The size, and not only the location of continents, must be considered. As long as continents are small and widely dispersed, precipitations can reach easily the inland areas. As a consequence, rainfall is not a limiting factor for runoff. Rather evaporation becomes the limiting factor, especially if the continents are located in the tropics, as already noticed by Barron et al. (1989). In this specific configuration (small continents and low latitude averaged location), runoff stays low, being constrained by the intense evaporation and responds weakly to global warming induced by a given CO_2 rise. Conversely, if small continents are moving poleward, the associated decreasing temperature induces a decrease in evaporation, possibly promoting runoff, and hence CO_2 consumption by weathering. Once cratons are aggregated into large continents (such as the Pangea), inland rainfall is strongly reduced and becomes limiting for runoff, instead of evaporation. Runoff is kept at low levels but its response to global warming events becomes stronger, forced by the rise in rainfall. Otto-Bliesner (1995) evidences the complex interplay between latitudinal location of the continents and their size. She emphasizes that the sensitivity of runoff to global warming might be very small for particular continental configuration, namely for small continents located at low latitudes.

In 1991, Bluth and Kump paved the way to a spatially-resolved calculation of the past weathering rates by defining lithological maps for the geological past, with a resolution of 2°long × 2°lat. Using their Quaternary map, they first calculated the weathering rates and fluxes during the last glacial maximum (Gibbs and Kump, 1994). In this study, the authors emphasize the importance of accounting for the spatial distribution in runoff when reconstructing past weathering rates. Going then into the distant past, Gibbs et al. (1999) explicitly calculate the spatial distribution of the weathering rates of the continents over the last 250 Myr, using a 3D general circulation model for the atmosphere (GENESIS v1.02). The solar constant is kept at present day value and two sets of simulation are performed: one assuming a constant CO_2 level of 340 ppmv, and one at high CO_2 taken from simulations of the GEOCARB geochemical model (Berner, 1991). The continental runoff and temperature maps generated by the climate model for each time slice are used to calculate weathering rate for each of the 5 main lithological types (granite, basalt, carbonate, shale and sandstone). The fundamental conclusion of their work is that the overarching effect of continental collision and accretion is increased continentality (first identified by Kutzbach and Gallimore in 1989 for Pangean climates), which inhibits runoff and hence CO_2 consumption by silicate weathering. Times of super continent assembly should correspond to periods of high pCO_2, keeping all other parameters constant, such as the solid Earth degassing rate. Overall, chemical erosion is limited by the supply of water, which depends heavily on the palaeogeographical setting (super continent versus dispersed continents).

4. Coupling climate and carbon cycle

4.1. GEOCARB and similar models

All the above models (Marshall et al., 1988; Worsley and Kidder, 1991; Otto-Bliesner, 1995; Gibbs et al., 1999) give insights into the role of palaeogeography on silicate weathering without computing the CO_2 levels. This has been achieved by another type of model, the most cited being the GEOCARB model (versions I, II, III, and GEOCARBSULF; Berner, 2004, 2006). The models of the GEOCARB family compute the Phanerozoic atmospheric CO_2 through the solving of a set of equations, the most important being the carbon and alkalinity balance at the million year timescale, the carbon isotopic balance, the expression of silicate weathering as a function of environmental parameters such as the vegetation cover, physical erosion, and climate. The GEOCARB models being zero-dimensional, they only calculate a mean global temperature and runoff, without resolving the spatial distribution of the continental climate. In details, mean global runoff and temperature are calculated through the solving of two equations.

$$T_{glob} = T_0 + \Gamma \log(pCO_2) - W_s \frac{t}{570} + geog \quad (1)$$

$$R_{glob} = R_0 \left[1 + Y\left(T_{glob} - T_0\right)\right] f_D \quad (2)$$

In Eq. (1), T_0 is the present day global air temperature, Γ is a parameter fixing the sensitivity of temperature to CO_2 variations, pCO_2 is the atmospheric CO_2 partial pressure in PAL (Pre-industrial Atmospheric Level), W_s is the rate of temperature rise with time related to the increase in the solar constant. In this equation, two parameters are functions of the continental configuration. Firstly the factor *geog* is the correction applied to the calculation of T_{glob} in response to continental drift. This factor is taken from Otto-Bliesner (1995) and is equal to the mean air continental temperature anomaly (relative to the present day control run) for each geological time slice from the simulations described above ($pCO_2 = 1$ PAL, no mountains, present day solar constant). Second, Γ is the sensitivity of the air temperature to a CO_2 change for a given palaeogeography. Two values are adopted for the Phanerozoic: 3.3 for a priori warm climate modes (before 340 Ma, and between 260 and 40 Ma, corresponding to a global warming of 2.3 °C for a CO_2 doubling), and 4 for other periods assumed to be cold (2.8 °C warming for a CO_2 doubling). Those values come from GCM simulation at various CO_2 levels for the present day (assumed to be representative of the Phanerozoic cold modes) and the mid-Cretaceous (a model for warm modes) configurations (Hay et al., 1999; Kothavala et al., 1999). Recently, Park and Royer (2011) have revisited the value of the Γ parameter, increasing it strongly during the cold phase of the Phanerozoic (Γ=8.7 to 11.5, corresponding to a 6 to 8 °C warming for a CO_2 doubling). Whatever the absolute values, it is important to note that the values of *geog* and Γ are fixed prior to the GEOCARB simulations, and do not evolve as a function of the calculated atmospheric CO_2. They are thus forcing functions and may introduce some inconsistency in the model: for instance low calculated CO_2 may be coeval with low climate sensitivity Γ.

In Eq. (2), R_0 is the present day runoff, *Y* is a factor fixing the sensitivity of runoff to air temperature, and f_D is an additional correction applied to runoff in response to continental drift. *Y* is fixed to 0.025 during warm climate modes (before 340 Ma and between 260 and 40 Ma, corresponding to a 2.5% increase in global runoff for a global mean temperature rise of 1 °C, Kothavala et al., 1999), and to 0.045 for other periods corresponding to a 4.5% increase in global runoff for a global mean temperature rise of 1 °C (Hay et al., 1999), a value close to the 4% estimated for the present day world by Labat et al. (2004) from river runoff time series. f_D is taken from the Otto-Bliesner (1995) simulations and is equal to the runoff calculated for each timeslice assuming 1 PAL of CO_2, flat continents and a present-day solar constant divided by the present day runoff.

These two equations are critical. They define how these key parameters of the silicate weathering rate (temperature and runoff) evolve through Phanerozoic times in response to CO_2 and tectonic changes. Being zero-dimensional models, GEOCARB and GEOCARBSULF cannot explicitly account for the role of palaeogeography, despite its potential key role (Marshall et al., 1988; Worsley and Kidder, 1991; Otto-Bliesner, 1995). To overcome this critical limitation, the GEOCARB family of models includes several parameters derived from climate evolution studies (such as Otto-Bliesner, 1995), in such a way that the role palaeogeography is somehow included (Γ, *geog*, f_D, and *Y* to some extent). The benefit of the use of these parameters to calculate temperature and runoff instead of physically resolving the role of palaeogeography is the gain in time-computation. But the risk of introducing inconsistencies, or overestimating or underestimating the role of palaeogeography cannot be neglected.

4.2. Sensitivity of the GEOCARB family of models to palaeogeography

Most of the other published models calculating past CO_2 levels use the same formalism for the impact of palaeogeography on climate (Wallmann, 2001, 2004) or simply neglect the potential effect of continental drift (Bergman et al., 2004).

What is the contribution of palaeogeography to the Phanerozoic CO_2 signal calculated by the GEOCARB type of models? To answer this question, we built a simplified version of the GEOCARB model (see Appendix A). When running the reference simulation for 21 time slices spanning the Phanerozoic, our simplified model version reproduces the general features of the GEOCARB family of models (Fig. 1): high CO_2 levels in the early Paleozoic, then a rapid decrease starting at 400 Ma. Low levels are predicted between 350 and 260 Ma. Then CO_2 rises again up to 10 times the present day values around 100 Ma. Finally, the late Cretaceous and Cenozoic are characterized by an overall CO_2 decrease towards the present day low level. We refer to this Phanerozoic trend as the "double-peak" feature in the following. Slight differences can be observed between this CO_2 curve and the latest GEOCARBSULF output, particularly higher levels around 300 and 100 Ma (Fig. 1). Indeed, we do not account for the role of the sulfur cycle, and silicate weathering laws and apparent activation energies for silicate dissolution are slightly different. Moreover, the temporal evolution of the continental area is taken from reconstructions by Blakey for the Paleozoic, Besse and Fluteau and Sewall for the Mesozoic and Cenozoic and differs thus slightly from the forcing used in GEOCARB. Finally, the time resolution of our simulations is much lower than Berner (2006). Anyway, our objective is not to reproduce precisely the GEOCARBSULF output, but rather to explore the contribution of the palaeogeographical setting to the "double-peak" evolution of atmospheric CO_2.

Then, we run our version of GEOCARB by fixing successively all parameters to their present day value except the parameters depending on the palaeogeography. At the end of this procedure (GLOBAL#4 simulation), only *geog*, f_D, Γ, and *Y* are allowed to fluctuate over the Phanerozoic. Solid Earth degassing rate is held constant and the organic carbon subcycle is assumed at steady state, which means that it does not impact the CO_2 level. A linear rise of the solar constant with time is still assumed. By doing so, we find that the double peak feature is mainly due to: (1) the impact of land plant evolution on continental weathering; (2) the variable volcanism; and (3) the fluctuations in the organic carbon cycle. When accounting only for the palaeogeographical parameters and the solar constant evolution, the major features of the GEOCARB CO_2 curve disappeared. The CO_2 displays now a general declining trend from about 6 PAL in the Cambrian to 1 PAL at present day, mainly forced by the rise in the solar constant. A sharp decrease around 460 Ma is predicted. This decline is related to the prediction of very high continental temperatures at 458 Ma in the simulations of Otto-Bliesner (1995), increasing sharply the *geog* factor, promoting silicate weathering, and thus forcing the predicted CO_2 to decrease to maintain the balance between silicate weathering and volcanic degassing. This warm climate is the result of the mean low latitude of the continental blocks during this period (Otto-Bliesner, 1995), and has been confirmed by more recent simulations (Nardin et al., 2011).

Corresponding calculated air temperatures are almost constant over the whole Phanerozoic, close to 14 °C (Fig. 2, GLOBAL#4 simulation), while runoff is predicted to be always larger than at present day, except around 250 Ma where it is very close to the present day value (Fig. 3, GLOBAL#4 simulation).

Overall, the sensitivity of the GEOCARB family of models to palaeogeography is low. The double-peak feature, which is the most visible feature of the CO_2 curve, is thus weakly dependent on the palaeogeographical setting.

4.3. The GEOCLIM model and the palaeogeography

4.3.1. Previous investigations with GEOCLIM

A new step in the carbon-climate modelling is achieved with Donnadieu et al. (2004) who introduce a new method to calculate the atmospheric CO_2 in the specific context of a supercontinent breakup at the end of the Proterozoic. The model they used, the first seed of the GEOCLIM model, couples a global carbon–alkalinity cycle to a climate model of intermediate complexity with a coarse spatial resolution (50°long × 10°lat). Continental weathering rates are resolved spatially, allowing to account explicitly for the continental configuration. Donnadieu et al. (2004) show that the Rodinia super continent,

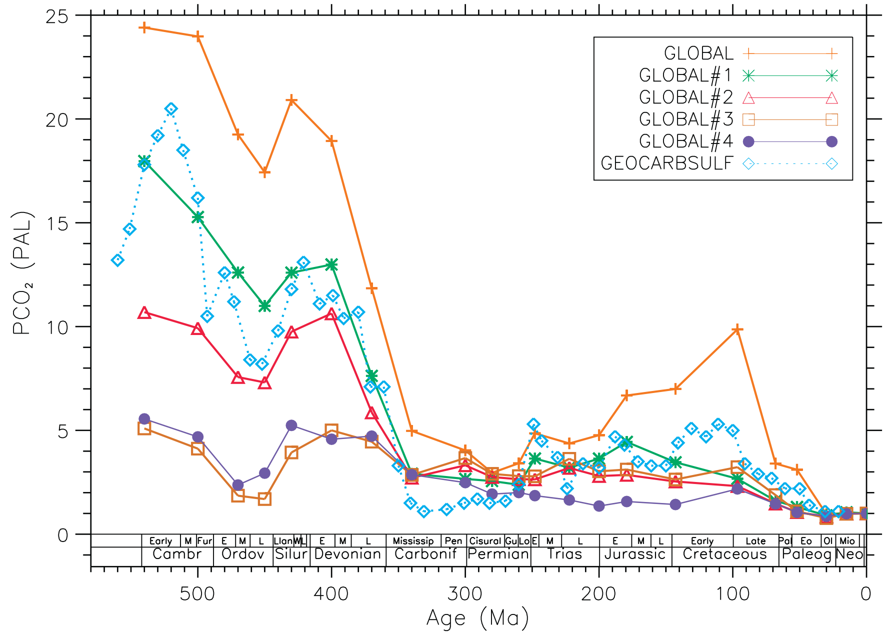

Fig. 1. Evolution of atmospheric CO_2 calculated by zero-dimensional models of the GEOCARB type. The curve labelled GEOCARBSULF is taken from Berner (2006). GLOBAL is the simplified version of the GEOCARB model built for the present purpose. The GLOBAL label reminds that this model includes a zero-dimensional description of the global climate (only global air temperature and runoff are calculated). GLOBAL#1 is the same simulation where the solid Earth degassing is held constant. GLOBAL#2 is the same as GLOBAL#1, but with the organic carbon subcycle always assumed at steady-state. GLOBAL#3 further assumes a constant impact of land plants on weathering (same as today). Finally GLOBAL#4 is the same than GLOBAL#3 but lithology and erosion rates are kept at their present day values. The only forcing functions of GLOBAL#4 are thus the rising solar constant and the palaeogeographical changes.

although located close to the equator, was characterized by low runoff, low weathering rates and hence high CO_2 (1800 ppmv). This is the result of the intense continentality, a finding in agreement with Gibbs et al. (1999). The subsequent breakup of the super continent into small pieces along the equator promotes the weatherability. The steady-state CO_2 level required for weathering to balance the solid Earth degassing equals about 500 ppmv in the dispersed continental configuration, corresponding to the onset of much colder climatic conditions compared to the super continent configuration (8 °C global cooling). When the weathering of freshly erupted basalts is accounted for, the breakup of the Rodinia might even be the ultimate cause for the onset of the Neoproterozoic snowball glaciations (Donnadieu et al., 2004).

The Donnadieu et al. (2004) model was a first 'low-resolution' synthesis of the ideas developed by Marshall et al. (1988) and Gibbs et al. (1999). Evolving from this first model, GEOCLIM is a numerical model of the next generation able to calculate the past geological CO_2 levels (Fig. 4). It couples an 11-box model of the carbon–alkalinity–phosphorus–oxygen cycles (Goddéris and Joachimski, 2004) to the 3-dimensional climate model FOAM (Jacob, 1997). At each time step of the calculation and for any CO_2 level, the output of the FOAM general circulation model is used to estimate the coeval air temperature and

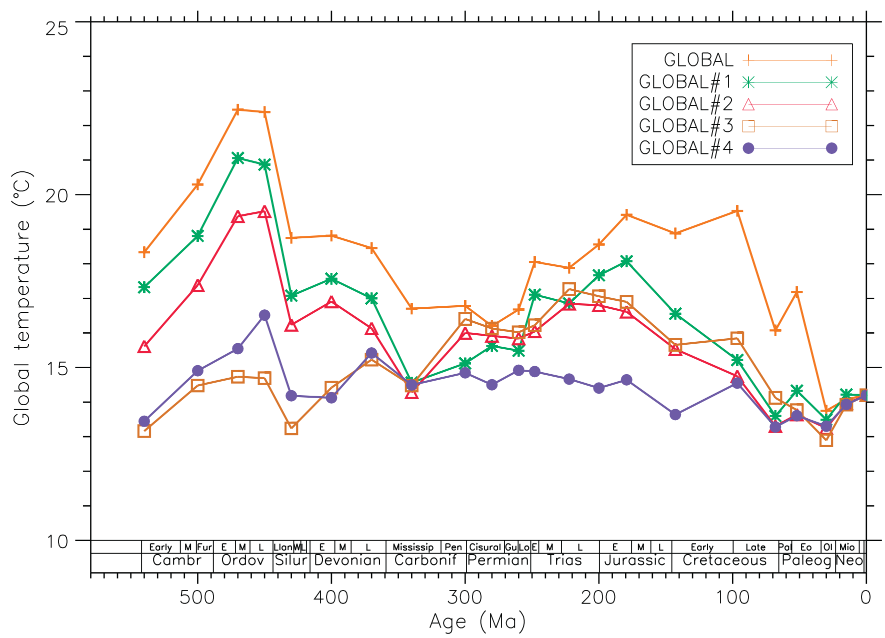

Fig. 2. Globally averaged surface temperature calculated in the various GEOCARB-type simulations described in Fig. 1. The curve GLOBAL#4 stands for the temperature change induced by the rising solar constant and the palaeogeographical changes solely.

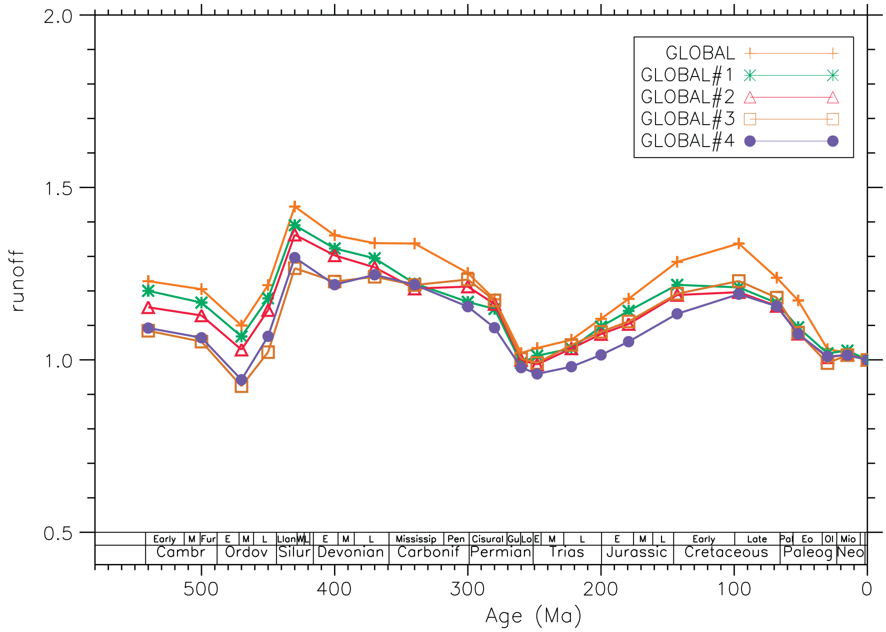

Fig. 3. Globally averaged continental runoff (normalized to present day) calculated in the various GEOCARB-type simulations described in Fig. 1. The curve GLOBAL#4 stands for the runoff change induced by the rising solar constant and the palaeogeographical changes solely.

continental runoff with a spatial resolution of 7.5°long × 4.5°lat. While GEOCARB is a zero-dimensional model (only able to calculate global averages) in which corrective factors have been introduced from several offline GCM simulations (Otto-Bliesner, 1995), GEOCLIM explicitly resolve the impact of palaeogeography coevally with the calculation of the atmospheric CO_2 level. It is important to note that the two methods (GEOCARB vs GEOCLIM) are fundamentally different. In GEOCLIM, temperature, runoff, and more generally the hydrologic cycle, are resolved physically by the GCM (conservation of mass and energy in the atmospheric system) in GEOCLIM while GEOCARB relies on parametric laws. Furthermore, in GEOCARB, *Y* and *geog* are fixed assuming a priori the timing of warm and cold climatic modes. In GEOCLIM, the sensitivity of runoff to temperature and of temperature to CO_2 is a function of the continental configuration, and fluctuates over the Phanerozoic. Furthermore, GEOCLIM allows the discrimination between continental and global temperatures.

In the following discussion, the continental runoff and its dependency on the continental configuration will be a key parameter. The mean global runoff calculated by the FOAM GCM for the present day configuration equals 330 mm/yr, in close agreement with the observations (313 mm/yr, UNH/GRDC database). Furthermore, the FOAM GCM is able to reproduce the observed spatial distribution of the continental

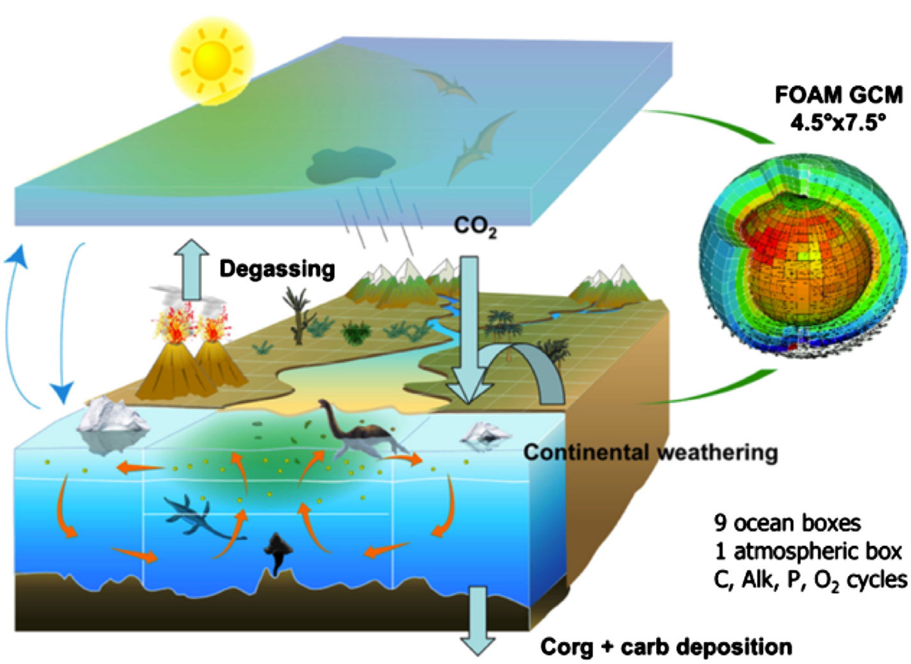

Fig. 4. Schematic description of the GEOCLIM model. A box model describing the geological cycles of carbon and alkalinity is coupled to the 3D FOAM climate model. Temperature and runoff calculated by the climate model are used to estimate the continental weathering fluxes, including the CO_2 consumption by silicate weathering at each time step with a spatial resolution of 7.5° long × 4.5° lat. For each continental configuration spanning the whole Phanerozoic, the GEOCLIM model is run until a steady-state is reached. Each continental configuration can thus be associated to a calculated CO_2 level, depending on other parameters such as the solid Earth degassing rate.

runoff, the main discrepancy between the model and the data being a slight underestimation of the runoff for the northern part of South America by the model (Fig. 5). Anyway, in terms of runoff predictions, the GEOCLIM model appears to be quite robust.

Several studies have been undergone with the GEOCLIM model, keeping all parameters constant but the continental configuration and the slowly rising solar constant. Nardin et al. (2011) demonstrate that the drifting of the Baltica, Laurentia, and Siberia across the equatorial belt promotes continental weathering. As a consequence, CO_2 goes down since the Middle Ordovician from about 5600 ppmv to 2800 ppmv to maintain the balance between silicate weathering and the solid Earth degassing. About 70% of this CO_2 drop is fully related to the drifting of continents, confirming the key role played by palaeogeography. The remaining 30% are due to enhanced young volcanic rock weathering (Nardin et al., 2011). The lowest CO_2 level is reached at the end of the Ordovician/beginning of the Silurian. Then CO_2 rises again in excess of 6000 ppmv in the Early Devonian because of the accretion of the Baltica and Laurentia. The continental configuration thus promotes the onset of colder climatic conditions from the Middle Ordovician to the Early Silurian, in agreement with many geological indicators (Nardin et al., 2011).

Still using the GEOCLIM model, Donnadieu et al. (2006) and Goddéris et al. (2008a) have shown that a <10° northward drift of the Pangea supercontinent from the Carnian to the Rhetian induces a drastic decrease in atmospheric CO_2 from about 3000 ppmv down to 800 ppmv in the late Triassic. This cooling was related to the specific shape of Pangea and does not involve any continental break up process. The continental surface located in the tropical humid zone is almost doubled when Pangea drifts 10° north. Silicate weathering is promoted and CO_2 goes down to maintain the carbon balance. This is in agreement with sedimentary $\delta^{18}O$ data which predicts a global cooling of 3.7 °C, closely matching the 3.3 °C calculated by GEOCLIM. When accounting for the continental drift, what matters is not only the latitudinal location of the continents, but also their size and our ability to reconstruct accurately the past water cycle.

An improvement of the GEOCLIM model is the coupling with a global dynamic vegetation model. Vegetation, which responds to the continental drift, can change the albedo of the continental surfaces, the cycling of water through enhanced evapotranspiration, and the roughness of the continents. All these factors may change the temperature and precipitation patterns, and hence the weathering rates and the calculated CO_2 level. Le Hir et al. (2011) and Donnadieu et al. (2009) have both demonstrated that the key climatic effect of land plants with land plants is to modify the surface albedo. For instance, in the Devonian, the colonization of continents by plants decreases the albedo, resulting in a global warming trend. This global warming promotes CO_2 consumption by continental weathering, and CO_2 is forced to decline to maintain the carbon cycle balance (Le Hir et al., 2011).

4.3.2. Exploring the impact of palaeogeography on the Phanerozoic CO_2 level

GEOCLIM is run for the 21 time slices covering the Phanerozoic. The model is run until steady-state is reached, meaning that continental silicate weathering matches the prescribed solid Earth degassing. We calculate the corresponding atmospheric CO_2 and climate for each of the 21 time slices. The simulations account for dynamic vegetation by coupling the FOAM GCM to the LPJ biospheric model (Sitch et al., 2003; Donnadieu et al., 2009). The FOAM GCM is used in mixed-layer mode, i.e., the atmospheric model is linked to a 50-metre mixed-layer ocean, which parameterizes heat transport through diffusion. Ocean surface temperatures are thus calculated, but not the oceanic circulation. A

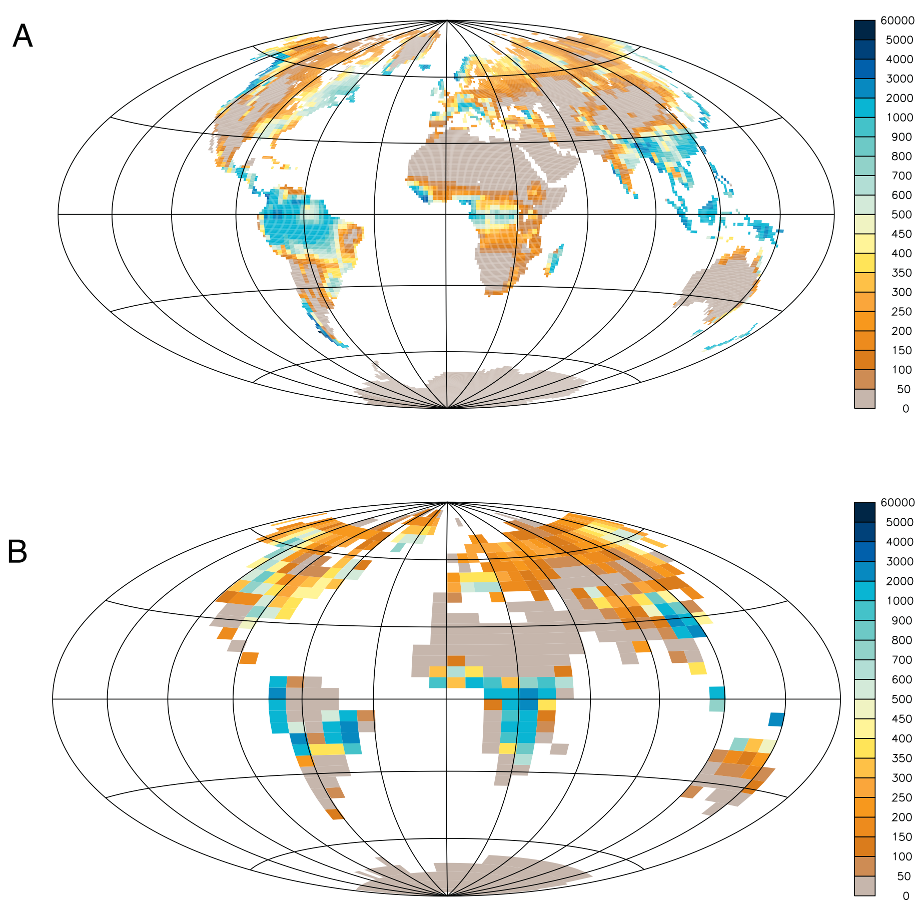

Fig. 5. (A) Present day observed distribution of the mean annual continental runoff (mm/yr) (UNH/GRDC: Composite runoff fields V1, 0.5°lat × 0.5°long). (B) Present day distribution of the mean annual runoff (mm/yr) calculated by the FOAM GCM under 280 ppmv CO_2.

first set of simulations is run assuming the palaeogeography and the increase in the solar constant as the unique forcing functions (GEOCLIMtec simulations). The second set of simulations accounts additionally for the role of land plants on weathering, for a variable degassing, and for imbalances in the organic carbon subcycles, all these forcing functions coming from the GEOCARB model and added in GEOCLIM (GEOCLIM simulations).

Accounting only for palaeogeography and the linear rise in the solar constant, the GEOCLIM CO_2 curve (GEOCLIMtec, Fig. 6) differs considerably from the CO_2 history produced by the GEOCARB family of models. A huge drop in CO_2 is predicted during the Ordovician and an ample CO_2 peak is calculated between 280–220 Ma. CO_2 levels are much higher before the mid-Cretaceous. Very low CO_2 levels dominate after 130 Ma (Fig. 6). The palaeogeographical imprint on the CO_2 history is major, contrasting with the GEOCARB-type simulation in which the palaeogeography has only a second order impact on the CO_2 level. High CO_2 levels at the beginning of the Phanerozoic are forced by the polar location of most of the continents. As identified by Nardin et al. (2011), atmospheric CO_2 then decreases due to the northward drift of Laurentia, Baltica and Siberia. This period of change is followed by a period of relatively constant CO_2 levels, suggesting that the palaeogeographical configuration has played a little role in the onset of the Permo-Carboniferous glacial episode. The end Permian to middle Triassic period is characterized by very high CO_2 levels, promoted by the assembly of Pangea resulting in a dry continental climate. CO_2 abruptly declines in the latest Triassic in response to a northward drift of Pangea (Donnadieu et al., 2006; Goddéris et al., 2008b). A striking feature is the persistent low CO_2 levels predicted for the Jurassic and Cretaceous. If palaeogeography is the only forcing factor, the post-Triassic dispersed continental configuration should favour the onset of a globally cold climatic mode since the Jurassic. The end-Cretaceous and beginning of the Cenozoic are even marked by CO_2 levels below the present day value.

Part of the GEOCLIM CO_2 signal is forced by the response of runoff to palaeogeography. The amplitude of change in runoff over the Phanerozoic is larger in GEOCLIM (both for GEOCLIM and GEOCLIMtec simulations) and in the simulations performed by Gibbs et al. (1999) compared to the simulations of Otto-Bliesner (1995) (Fig. 7). Furthermore, in Otto-Bliesner (1995), the runoff is generally higher than at present day, except for the Permo-Triassic period where it is close to the present day value. This explains partly why the CO_2 levels calculated by GEOCARB (GLOBAL#4 run, only solar constant and palaeogeography changing) are much lower than in the GEOCLIM run (GEOCLIM run) (Fig. 8). Consistently with Gibbs et al. (1999), the GEOCLIM runoff is conversely generally lower than at present day, at least before the mid-Cretaceous. Consequently, the GEOCLIM CO_2 is higher, because weathering is partly inhibited. The dry peak of the Permo-Triassic is also much more ample in more recent GCM simulations than in Otto-Bliesner (1995), explaining the existence of a calculated CO_2 peak (global runoff is below 60% of the present day value).

It is not easy to identify a unique cause for the calculated CO_2 features by GEOCLIM. Considering the GEOCLIMtec simulation, where continental drift is the sole forcing function, a plot of the calculated atmospheric CO_2 versus the total continental surface allows us to identify at least three groups of points (Fig. 9A). The first group has a standard behaviour ("standard" pool): atmospheric CO_2 decreases gently with the continental surface. More continental surface implies more silicate weathering, and hence less CO_2. Then the second group displays a range of high CO_2 values despite a roughly constant continental surface. This is the case for the 260–220 Ma time window and for the early Phanerozoic. These points define the "warm" Phanerozoic pool. Finally, a last group displays very low CO_2 levels whatever the total continental size. This group is identified here as the "cold" pool and covers the end of the Phanerozoic, starting from the middle Cretaceous.

The existence of a "warm" pool is linked to the aridity developing during the assembly of the Pangea super continent (Fig. 9B), but not only. This "warm" pool is also the consequence of a weak fraction of emerged continents in the warm and humid equatorial zone (10°S–10°N), where modelled weathering fluxes are maximized (Fig. 9C). This is the case during the Pangea phase and also during the early Phanerozoic. Consequently, the global weathering rates are inhibited and CO_2 rises. The existence

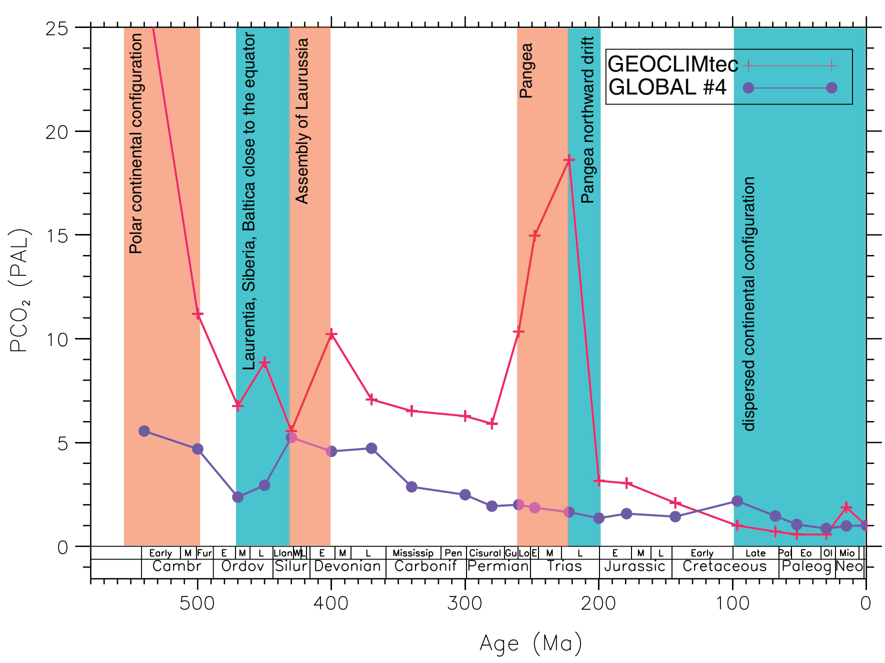

Fig. 6. Calculated evolution of the atmospheric CO_2 level over the Phanerozoic. The curve GLOBAL#4 stands for the CO_2 evolution induced by the rising solar constant and the palaeogeographical changes solely in a GEOCARB-type simulation. Similarly, the GEOCLIMtec curve is the output of the GEOCLIM model when the only forcings are the solar constant and the palaeogeographical changes. Accounting for the spatial distribution of the rainfall and temperature (and hence of weathering) in GEOCLIM impacts strongly on the calculated CO_2 history, suggesting that the palaeogeographical setting is a first order forcing function of the Phanerozoic climatic evolution. Red boxes figurate periods where the palaeogeography acts as a warming forcing function, promoting high CO_2 levels. Conversely, blue boxes stand for period where the continental configuration acts as a cooling forcing function. The ultimate causes for these trends are written in each box.

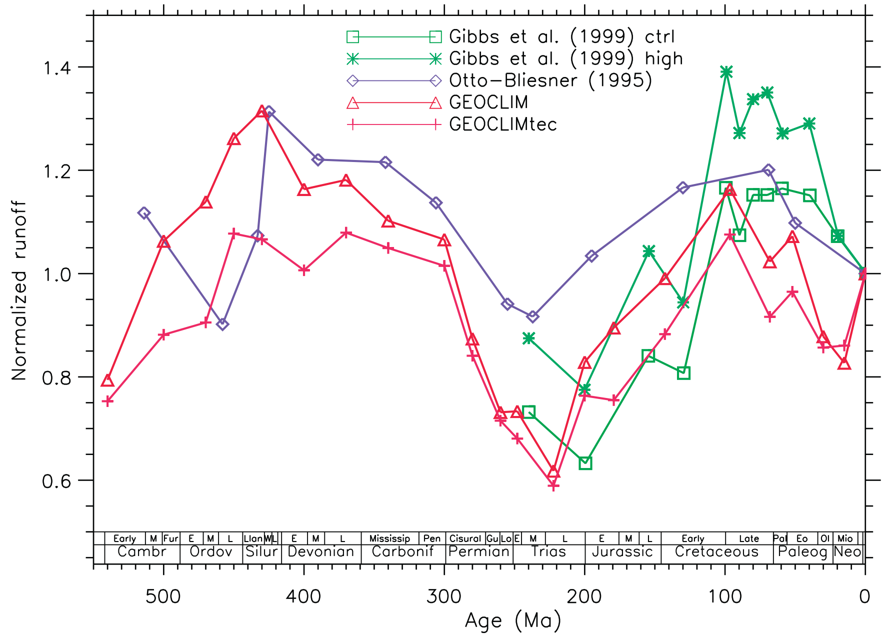

Fig. 7. Time evolution of the global runoff calculated by 3D climate models. The Gibbs et al. (1999) ctrl simulation has been performed assuming a constant 340 ppm of CO_2 and moving continents. The Gibbs et al. (1999) high simulation was performed assuming variable CO_2 taken from GEOCARB. The Otto-Bliesner (1995) simulation was performed assuming 280 ppm of CO_2. Finally, GEOCLIM and GEOCLIMtec are respectively the output of the full GEOCLIM simulation, and of a GEOCLIM simulation where the only forcings are the rising solar constant and the palaeogeography.

of a "cold" pool since the middle Cretaceous is linked to the ability of the Earth system to sustain an intense runoff over large continental surfaces (Fig. 9B). This is true even if the continental surface in the equatorial area stays generally below the mean Phanerozoic value. Runoff is no more dependent on the emerged continental size and stays at a high value. Since about 90 Ma, the continental configuration (highly dispersed continents, with only small polar continents) maintains high runoff and hence high CO_2 consumption by silicate weathering, forcing the Earth system into a cold state.

Consistently with Otto-Bliesner (1995), we found a dependence of runoff on the averaged latitudinal location of the continents (Pearson correlation coefficient 0.60 in the GEOCLIMtec simulation) (Fig. 10A). The highest runoff rates coincide with the highest averaged latitudinal location. At first glance, this result is in agreement with previous calculations performed with generic continental configurations in which an equatorial "ring" world (extending from 17°S to 17°N) was characterized by low global runoff (Hay et al., 1990). Indeed, large continental surfaces located along the equator may inhibit the evaporation in the tropical convergence zone, decreasing the global runoff. However, in our simulations, the increase in global runoff with the mean latitude of the continents can be explained by simple "geometric" consideration. Whatever the considered period, runoff is always minimal between

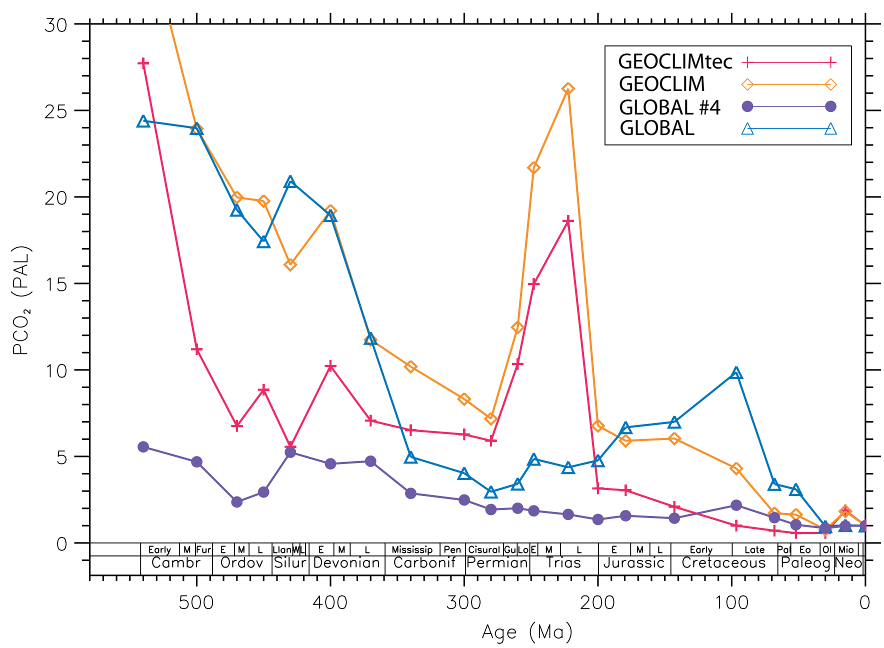

Fig. 8. Phanerozoic CO_2 history calculated by the zero-dimensional GEOCARB-type model and by the three-dimensional GEOCLIM model. GLOBAL#4 and GEOCLIMtec are simulations where the only forcings are the rising solar constant and the palaeogeography.

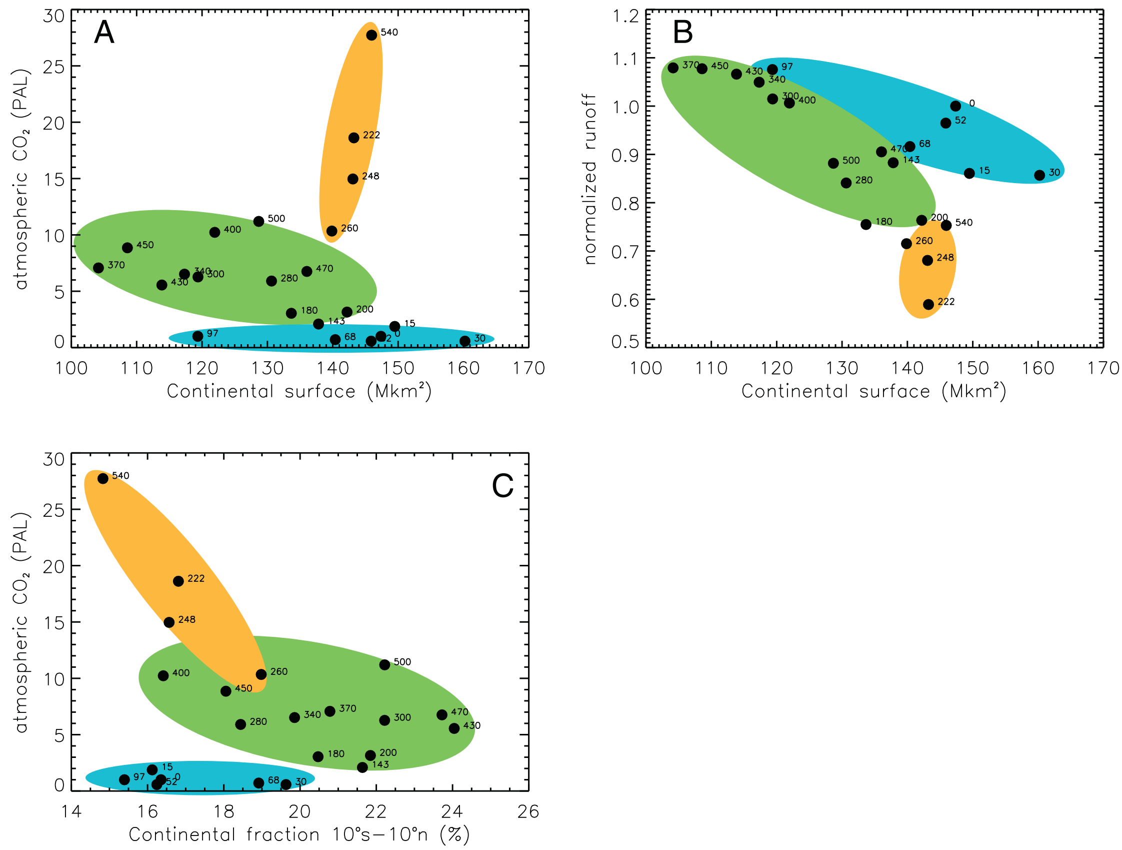

Fig. 9. (A) Atmospheric CO_2 versus the total emerged continental surface in the GEOCLIMtec simulation. The black numbers stand for the age of each dot, in million years. (B) Global continental runoff normalized to its present day value versus the total emerged continental surface in the GEOCLIMtec simulation. (C) Atmospheric CO_2 versus the percentage of the total continental surface located in the 10°S–10°N latitude band in the GEOCLIMtec simulation. The orange envelope represents the "warm" pool, the blue envelop the "cold" pool and the green envelop the "standard" pool. Each of these pools is exclusively forced by the palaeogeographical forcing.

roughly 15° and 35° latitude, because this zone coincides with the subsidence of the Hadley cell. The width of this low rainfall/high evaporation area is somehow dependent on the continental configuration, but not critically. As a consequence, when the continental fraction increases in the 15°–35° zone, we might expect that the global runoff will decrease simply because the continental area exposed to very low rainfall increases. This "geometric" effect is illustrated on Fig. 10B, where the global runoff is negatively correlated to the continental fraction in the 15°–35° latitude band (Pearson correlation coefficient −0.84). The positive correlation between the averaged latitudinal position of the continents and the runoff arises from the "fortuite" existence of a negative correlation between the averaged latitudinal location of the continents and the continental fraction between 15°–35° latitude (Fig. 10C).

It is worth mentioning that both model (GEOCLIM and GEOCARB) have a rather different behaviour with respect to the Permo-Triassic dry event. Although less pronounced than in GEOCLIMtec, the runoff calculated in GLOBAL#4 simulation also displays a decrease around 250 Ma (both GEOCLIM and GEOCARB-style simulations being performed with the palaeogeographical forcing only). Despite this runoff decrease by a about 30% from the Carboniferous to the Permo-Triassic, calculated atmospheric CO_2 in GLOBAL#4 does not display any rise in CO_2 at the same time, while GEOCLIMtec peaks at almost 20 times the present day level with a 40% decrease in runoff (Fig. 6). The reason for these distinctive behaviours is rooted in the parametric expression of silicate weathering as a function of runoff. In GEOCLIMtec, silicate weathering is linearly dependent on runoff, as suggested by recent compilations of field studies of silicate weathering (Dessert et al., 2003; Oliva et al., 2003). In GLOBAL#4, silicate weathering linearly increases with the runoff to the power 0.65, to account for dilution effects (Berner, 1994, 2004). This dilution effect is introduced to fit the database of Dunne (1978) for Kenyan rivers and of Peters (1984) for rivers of the United States. As a result, the response of weathering to runoff changes is seriously damped in GLOBAL#4, explaining why there is no CO_2 peak during the Permo-Triassic low runoff interval. This illustrates that the superposition of numerous parametric laws of various origin to describe the climate/carbon system, may finally hide the role of the palaeogeography on the atmospheric CO_2 level. This result also stresses that runoff and its impact on the CO_2 consumption by rock weathering must be modelled accurately. The use of a GCM to simulate the water cycle in GEOCLIM is a major improvement compared to zero-dimensional models, but we are still dependent on the reliability of parametric laws to describe silicate weathering. Because these parametric laws are fitting functions of the field data (for instance, HCO_3^- fluxes at the watershed outlet versus mean annual runoff, Dessert et al., 2003), the most complete databases must be used. Both the Oliva et al. (2003) and Dessert et al. (2003) databases cover a wide variety of climatic environments. But the spreading of their data points, particularly important for granitic environments, still questions the reliability of parametric laws. This is an inherent limitation of the method.

4.3.3. Role of small continents

The drastic decrease in atmospheric CO_2 in the Rhetian stage of the Triassic is mainly linked to the Northward drift of Pangea (Goddéris et al., 2008a). But there is an additional contributor to this sharp cooling trend. South China is located in the humid tropical area at the same time. In the simulation where the continental configuration is the only forcing (GEOCLIMtec), silicate weathering in South China is accounting for 17% of the global CO_2 consumption by weathering (Fig. 11). If CO_2 consumption by silicate weathering in South China during the Rhetian is removed from the simulation, atmospheric CO_2 would rise from

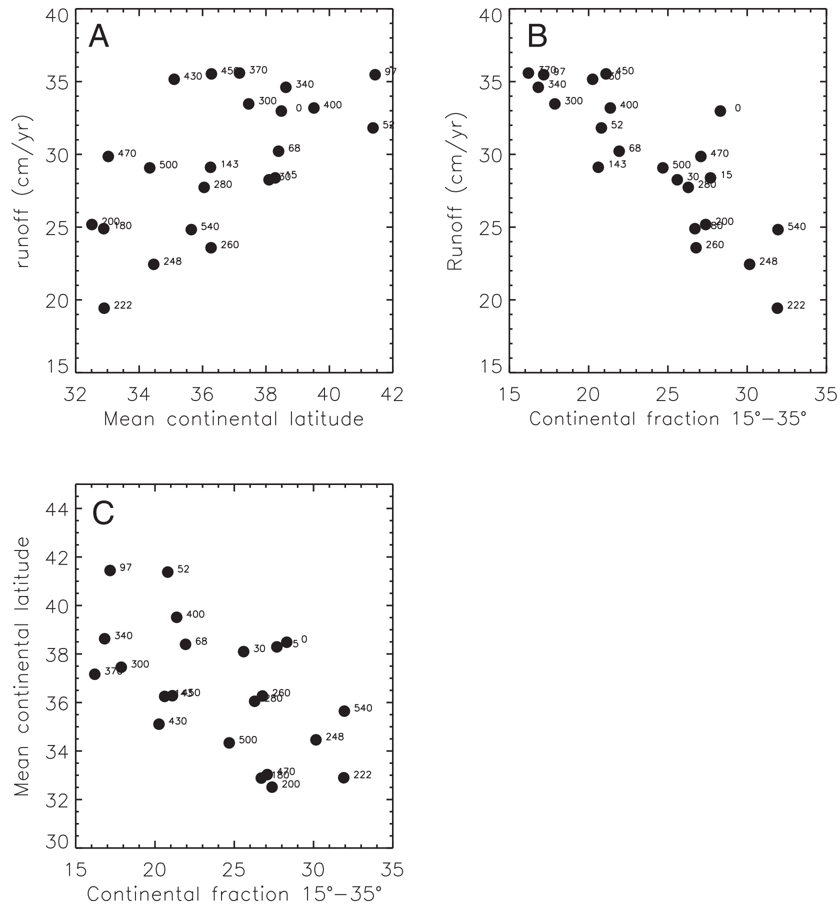

Fig. 10. (A) Global runoff as a function of the averaged latitudinal positions of the continents. (B) Global runoff as a function of the continental fraction within the latitude band extending from 15° to 35° latitude, in both hemisphere. (C) Averaged continental latitude of the continents as a function the continental fraction within the latitude band extending from 15° to 35° latitude, in both hemisphere. Runoff output from the GEOCLIMtec simulation. The black numbers stand for the age of each dot, in million years.

3.15 PAL (880 ppmv) to 5 PAL (1400 ppmv). Because of the logarithmic dependence of temperature on CO_2, the subsequent global warming of the Rhetian stage is significant (3 °C continental warming). This sensitivity test emphasized the potential key role played by small continents on the global carbon cycle, especially when they are located in the humid tropical belt (Oglesby and Park, 1989). This result somehow contradicts Otto-Bliesner (1995), who found that small continental blocks located in the tropical area might be characterized by low runoff, because evapotranspiration exceeds rainfall. If this was the case, small continental blocks should have only a minor effect on the long-term climate evolution. Conversely, our simulations show that small tropical continents might contribute significantly to the global CO_2 sink by weathering.

4.3.4. Climate sensitivity and palaeogeographical configuration

In the recent years, geological climate reconstructions have been used to assess the climate sensitivity to atmospheric CO_2 changes (Berner and Kothavala, 2001; Lunt et al., 2010; Park and Royer, 2011).

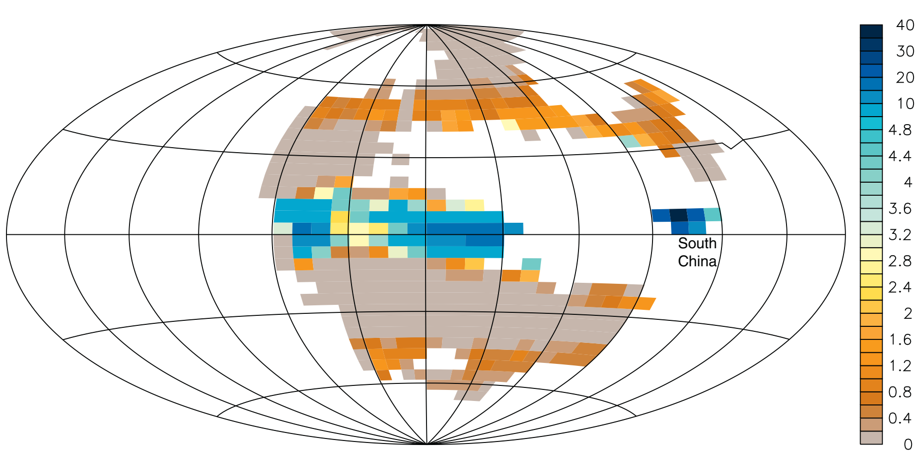

Fig. 11. CO_2 consumption by silicate weathering (10^{10} mol CO_2/yr) calculated for the Rhetian stage of the Triassic in the GEOCLIMtec simulation (pCO_2 = 5040 ppmv). South China crosses the equatorial humid belt and consumes 17% of the total CO_2 consumption.

We show here that climate sensitivity is substantially dependent on the palaeogeographical setting. Sensitivity is maximum during the Early Cambrian and during the Early to Middle Triassic period, exceeding 4 °C for a CO_2 doubling (Fig. 12). Except for this two particular geological period, the sensitivity is more less constant oscillating around 2.8 °C with a minimum at 2.4 °C at 90 Ma. What make the Early Cambrian and the Early to Middle Triassic so sensitive to a CO_2 doubling is not straightforward at first. It is important to note that these calculations of the climatic sensitivity account for fast and slow climatic feedbacks, excepting the role of the cryosphere, which is not modelled.

Climate sensitivity seems to respond partly to fluctuations in the continental surface as a function of latitude (Fig. 12). It roughly follows the evolution of the continental fraction between 60° and 30° latitude. Sensitivity increases when mid-latitudinal continental fraction rises, as it is the case in the Early Cambrian and during the Pangean period. But in details, things are trickier. The rapid increase in climate sensitivity from about 2.8 °C to more than 4 °C around 240 Ma is correlated to a strong decrease in the polar continental fraction. Also during the end Mesozoic and the Cenozoic, climate sensitivity decreases while the mid-latitude continental fraction stays roughly constant.

Runoff dependence on global temperature is also modulated by the palaeogeographical evolution. During most of the Paleozoic, from the Ordovician to the Mississippian, the runoff sensitivity to temperature is around 1.5 cm yr^{-1} K^{-1} (Fig. 13). Then it rapidly decreases with the assembly of Pangea, reaching 0.6 cm yr^{-1} K^{-1}. The Mesozoic is then characterized by a slow increase in runoff sensibility. Finally a major rise is predicted in the Miocene, where sensitivity goes back to Paleozoic values.

4.3.5. GEOCLIM output and proxy data

The objective of this contribution is to emphasize the key role played by the palaeogeographical configuration on the carbon cycle and on the climate evolution. At this stage, it is not easy to compare the GEOCLIM outputs with proxy data, because GEOCLIM has been designed to model mechanistically the role of palaeogeography, and does not aim at being a model of the complete carbon cycle. The other processes, including the role of the organic carbon cycle, of the vegetation on weathering, are either not modelled (GEOCLIMtec simulation), or modelled using simple parametric laws (GEOCLIM simulation). Nevertheless, some conclusions can be drawn from the comparison between proxies and model output.

Regarding the CO_2 level, the agreement between the GEOCLIM outputs and the CO_2 proxies is acceptable only for the last 200 Ma of the Earth's history, considering the large uncertainties inherent to the reconstructions (Fig. 14A). Between 200 and 400 Ma, GEOCLIM tends to overestimate the past CO_2 levels. This overestimation is particularly striking in the Triassic, where the supercontinental configuration forces GEOCLIM to very high CO_2 values. These high values are not recorded in the CO_2 proxies. Then, the 250–350 Ma period is characterized by very low proxy-based CO_2 values, around 500 ppmv, while GEOCLIM predicts values in excess of 1000 ppmv.

We also compare the mean annual sea surface temperature (SST) calculated by GEOCLIM within the 30°S–30°N latitudinal band to the temperature reconstructed using the tropical $\delta^{18}O$ data compiled by Prokoph et al. (2008) (Fig. 14B). To translate $\delta^{18}O$ into SST, we follow the method described in Veizer et al. (2000). $\delta^{18}O$ data have been detrended, using a root mean square linear fit. The detrending is applied to avoid unrealistically high temperatures mainly before 100 Ma. Then a running average has been calculated (20 Myr time window), including the calculation of the range containing 68% of the data points. Finally, we set the seawater $\delta^{18}O$ to −0.7 permil in the Cenomanian (a period where continental ice is assumed to be absent), and to 1 permil for the coldest stage of the Phanerozoic (most positive $\delta^{18}O$ detrended value). Because this represents a significant handling of the data, we express the result in terms of relative changes to the present day instead of absolute temperatures. Furthermore, the detrending procedure may hide real long-term climatic trends in the oxygen record. This is particularly critical for the last 100 Ma of Earth history, where the $\delta^{18}O$ signal provides a realistic temperature signal even without detrending (Donnadieu et al., 2009). For that reason, we also display on Fig. 14B the temperature calculated for the last 100 Ma without detrending of the oxygen isotopic data.

The spreading of the temperature values is quite large. The most striking feature is the existence of a warm peak in the Late-Permian/Early Triassic, in agreement with the GEOCLIM output (Goddéris et al., 2008a). A recent study has evidenced the

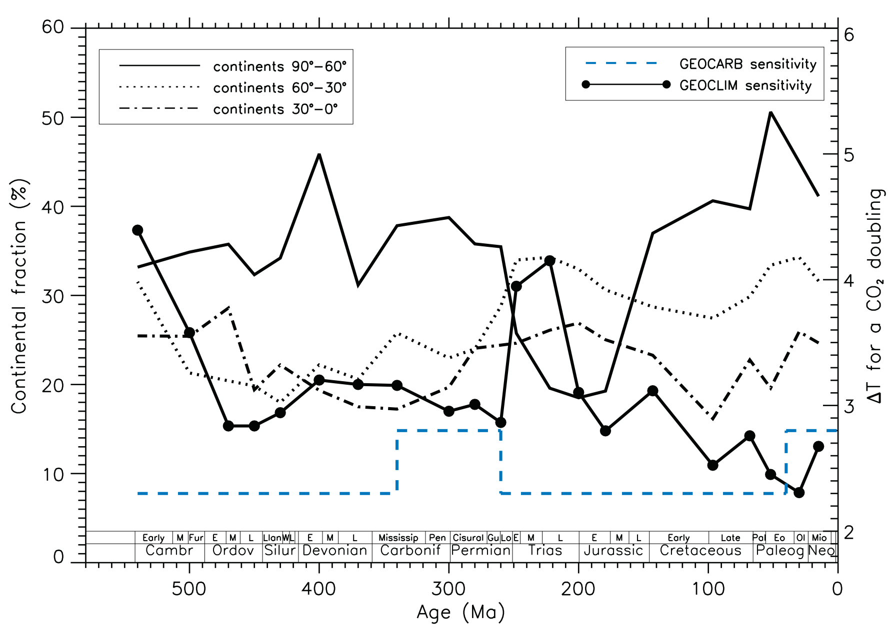

Fig. 12. Mean global temperature change for a CO_2 doubling as a function of age, calculated by the GEOCLIM model (solid line, scale on the right). The dotted blue line stands for the climate sensitivity prescribed in the GEOCARB model. Superimposed is the fraction of continents in various latitude bands (scale on the left).

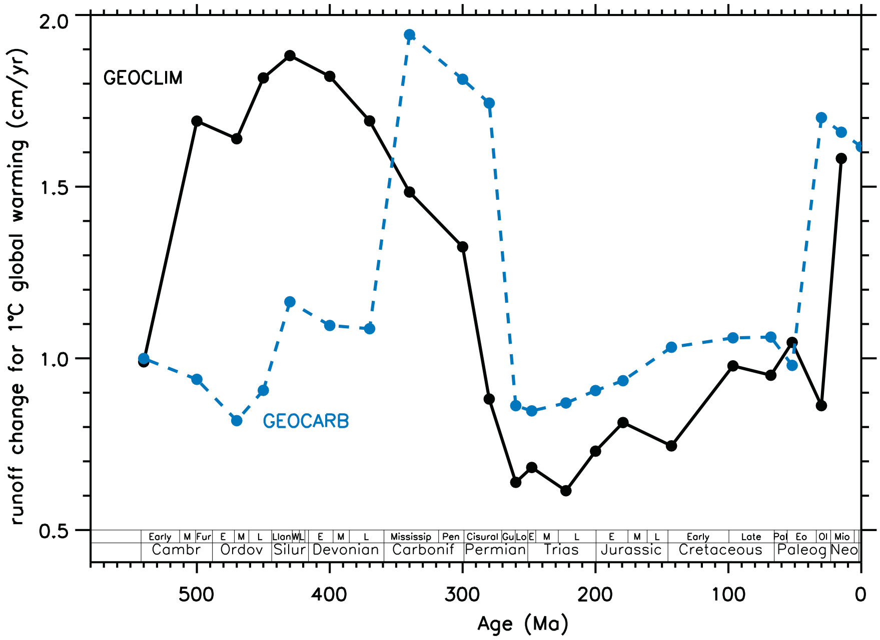

Fig. 13. Runoff change in cm/yr for an increase of 1 °C in the global temperature, calculated by the GEOCLIM model using a three-dimensional calculation of the water budget (solid line). The dotted blue line is the same calculated by the zero-dimensional GEOCARB model, which includes a parametric description of the water cycle.

persistence of very high temperatures in the tropical ocean over the whole Early Triassic, in excess of 35 °C (from conodont apatite $\delta^{18}O$) (Sun et al., 2012). As noted by the authors themselves, high temperatures for the 5 Myr of the Early Triassic require strong and persistent greenhouse conditions. These high temperature conditions may have lasted until the end of the Carnian stage, as inferred from brachiopod $\delta^{18}O$ (Korte et al., 2005). In the GEOCLIM simulations as well as in the Gibbs et al.'s (1999) study, these long lasting warm conditions can be sustained by the developing aridity, limiting the efficiency of continental weathering and promoting high CO_2 levels for millions of years. Conversely, the major temperature decrease occurring after 100 Ma and predicted by the model is not observed in the Prokoph et al. (2008) database when the oxygen isotopic data are detrended. But a cooling of about 5 °C in the tropical SST is reconstructed between 100 and 0 Ma when the oxygen isotopic data are not detrended, in close agreement with the model output (Donnadieu et al., 2009).

Several conclusions can be drawn. (1) When considering the early Triassic warm event, there is a clear disagreement between CO_2 proxy data and temperature data that we already note in a previous contribution (Goddéris et al., 2008a). A warm event seems to be associated to low CO_2 values. A coupled carbon-climate model like GEOCLIM, and even GEOCARB, cannot calculate at the same time high temperatures and low CO_2 levels. This apparent discrepancy between proxy-based temperatures and proxy-based CO_2 levels might be linked to the pretty large uncertainties inherent to the reconstruction methods, but this is beyond the scope of this study. (2) It appears that the palaeogeographical forcing is not responsible for the Permo-Carboniferous cool mode. (3) Reconstructing $\delta^{18}O$-based temperature curves for the whole Phanerozoic is not an easy task. However such composite regional or latitudinal curves would greatly help at validating spatially resolved numerical models. The various available databases are not really self-consistent. They mix phosphate and carbonate signals, the mathematical relationship between $\delta^{18}O$ and temperature fluctuates from one study to the other, and the data reflects the temperatures at various depths, depending on the organisms. Furthermore, the role of diagenesis is still not fully understood. Most of the geochemists now rely preferentially on phosphatic data for periods older than the Jurassic (Trotter et al., 2008; Joachimski et al., 2009), because phosphatic materials are much less sensitive to diagenesis than pre-Jurassic carbonates. But the question of the phosphate–water fractionation equation is still debated (Longinelli, 2013; Pucéat et al., 2013) and the question of the seawater $\delta^{18}O$ value is still critical. (4) Non-quantitative data should not be neglected when using spatially resolved models. Indeed, the GEOCLIM predictions of a very warm early Triassic are in agreement with sedimentological evidence for the deposition of redbeds and massive evaporites. The calculated sharp cooling during the latest Triassic coincides with major climatic changes impacting various regions of the world. Pedogenic and clay mineral data (smectite-rich redbeds) indicate a switch from arid/semi-arid conditions to year-round humid environments, a switch reproduced in our simulations (see Goddéris et al., 2008a, for references and more details). (5) The aim of the modelling efforts described in this work is not really to improve the results in terms of past atmospheric CO_2, but rather to improve the method by introducing physics in the numerical reconstructions of the past climate and carbon cycle. The GEOCARB model paved the way of the quantitative reconstructions of the Phanerozoic carbon cycle. The method used was mostly parametric. Now, recent studies relying on complex coupled climate-carbon numerical models have shown that the palaeogeographical factors have been underestimated. This review demonstrates that we now understand how and to what extent the palaeogeography really impacts the past carbon cycle and climate. This is a good starting point for revisiting the other processes at play, such as the links between weathering and land plants, between weathering and relief, among other parameters.

5. Conclusions

Two modelling pathways have been followed in the recent years for reconstructing the geological atmospheric CO_2 content. Zero-dimensional models (the so-called GEOCARB model family in this contribution) have been developed with the objective of accounting for as much processes as possible, such as the role of land plants or physical erosion on the weathering fluxes. Such models also include a simple parametric description of the climate response to the CO_2 content, the solar constant rise or the continental configuration on a global basis.

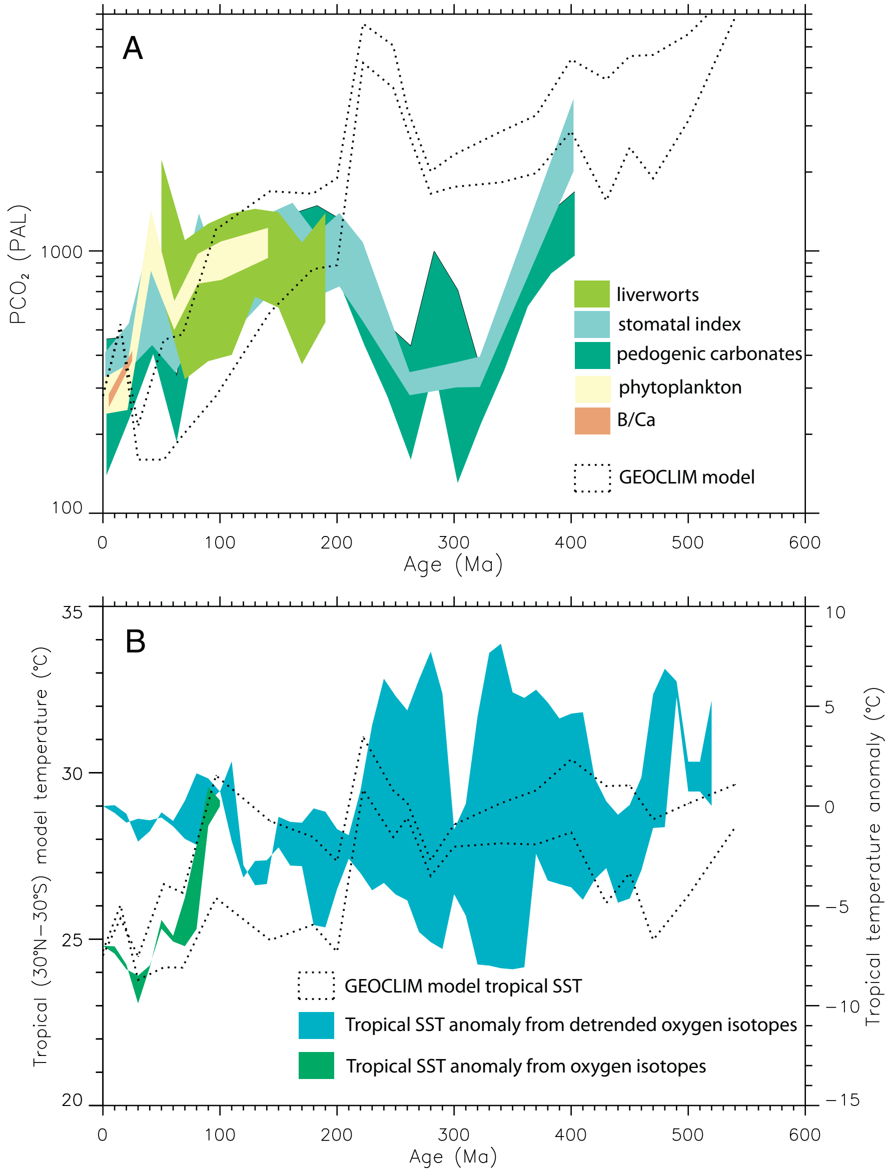

Fig. 14. (A) Comparison between the atmospheric CO_2 level calculated by the GEOCLIM model (dashed lines, the upper curve is the GEOCLIM simulation, the lower curve is the GEOCLIMtec simulation) and the CO_2 levels from various proxies (20 Myr moving averages). (B) The tropical sea surface temperature (averaged 30°S–30°N latitude) calculated by the GEOCLIM model (dashed lines, the upper curve is the GEOCLIM simulation, the lower curve is the GEOCLIMtec simulation, scale on the left). Also shown is the temperature anomaly for the same zonal area estimated from sedimentary carbonate $\delta^{18}O$ (scale on the right). See text for more details.

Those models gather information from other model runs or from field data and include those as parametric laws. Mixing several parametric descriptions from various origins is always at risk of losing the inherent physical behaviour of a complex system such as the Earth system. Nevertheless, these zero-dimensional models have the advantage of being integrative (many processes are considered), and fast in terms of computation time. They also generate past CO_2 reconstruction that can be compared to available data because they aim at accounting for all the factors controlling the evolution of the global carbon cycle.

Since the end of the 80s, physically-based models have also been developed to describe specifically the impact of the continental configuration of the global carbon cycle and climate geological evolution. Such models all include a spatially-resolved description of the climate system. This is the minimum requirement to consider physically the role of the palaeogeography on the global carbon cycle. Those models, one to three-dimensional, proceed from a different philosophy than the zero-dimensional method. The objective is no more to calculate a time evolution of the atmospheric CO_2 than can be compared readily to data, but to explore the sensitivity of the Earth system to one parameter, here the palaeogeographical configuration. These studies come to four major findings:

(1) The palaeogeography is a key controlling factor of the geological evolution of the atmospheric CO_2. By modulating runoff and air temperature above the continental surface, it strongly controls the CO_2 consumption by silicate rock weathering. Supercontinent and polar configurations favour the onset of very high CO_2 pressure in the atmosphere. Conversely, since about the mid-Cretaceous, the palaeogeography promotes the onset of low CO_2 and cold climatic conditions. The role of the palaeogeography has been largely underestimated in zero-dimensional models.

(2) Some of the key features of the Phanerozoic climate can be explained by the modulation of the carbon cycle by continental drift. This is the case for the early Paleozoic (Cambrian to early Silurian) global cooling, the warm and arid early Triassic, and the cold trend of the Cenozoic.
(3) Small continental blocks may contribute significantly to the total CO_2 consumption by weathering when entering the sub-tropical warm and humid zone.
(4) Climate sensitivity is substantially dependent on the continental configuration. As such, though the past might be the key to estimate the amplitude of the future climatic changes, it is important to understand the processes linked to the palaeogeography that have modulated the Earth climatic sensitivity in the past.

Instead of using parametric relationships to estimate global runoff and temperature, the models described in this review (Otto-Bliesner, 1995; Gibbs et al., 1999; Donnadieu et al., 2004, 2006; Goddéris et al., 2008b; Nardin et al., 2011; Lefebvre et al., 2013) resolve spatially the climatic conditions using process-based climate models. Such models implement the energy budget of the Earth atmosphere and oceans and a physical description of the water cycle. Those models are not yet able to propose a CO_2 history that can be compared to data. Indeed, despite a major improvement in the physical basis of the climate compared to parametric models, one major constraint remains the mathematical expression for CO_2 consumption by continental weathering. In Gibbs et al. (1999) and in the GEOCLIM model, continental weathering is calculated on each grid cell assuming runoff (and temperature) dependencies. But weathering is a complex process. It also depends on physical erosion, vegetation cover, soil thickness, and lithology. Regarding the vegetation cover, a few studies try to explore physically the climatic (Donnadieu et al., 2009; Le Hir et al., 2011) and geochemical (Taylor et al., 2012) impacts of land plants within the framework of a three-dimensional climate and geochemical modelling. All these studies show the critical role played by the land plant distribution on the weathering fluxes, and hence on the global carbon cycle. Regarding the thickness of the soil cover, protecting the bedrock from weathering when it becomes too thick, Goddéris et al. (2008b) have shown, using the GEOCLIM model that this effect has modulated the CO_2 levels. But the role of large mountain ranges on the global carbon cycle is still to be included in those spatially-resolved numerical model. Finally, although Gibbs et al. (1999) suggest that spatial variations in lithology account only for small fluctuations in the total chemical erosion rates, this was not the conclusion of Taylor et al. (2011, 2012), Nardin et al. (2011) and Kent and Muttoni (2013). Indeed these authors found that the bulk of weathering may come from basaltic hotspots.

In summary, it appears that we are slowly moving from simple parametric representations of the geological Earth system towards more comprehensive and physically-based descriptions. What we have learnt in the recent years is that the overall picture we had of the carbon cycle and climate evolution may drastically change when more complex models are used. Further works are clearly needed to account for a better representation of the weathering rates for instance, but they should include, as far as possible, a description of the physics at play.

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Appendix A

Basically, the calculation of the Phanerozoic atmospheric CO_2 pressure by the GEOCARB family of models relies upon a set of equations. Although the complete GEOCARB model is more complex (Berner, 2004), we only show here the key equations. The objective is to emphasize the role played by the palaeogeography on the climate evolution of the Earth, and not to reconstruct strictly the GEOCARB model.

(1) The carbon balance: given the short residence time of carbon in the ocean/atmosphere system compared to the geological timescale, the exospheric carbon cycle is assumed to evolve through a succession of steady states:

$$F_{vol} + F_{MOR} + F_{cw} + F_{ow} = F_{cd} + F_{od} \tag{A1}$$

where F_{vol} is the supply of CO_2 by subaerial volcanism, F_{MOR} is the degassing flux of CO_2 at mid-oceanic ridges, F_{cw} is the carbon flux released by continental carbonate rock dissolution, and F_{ow} is the carbon flux released to the river system by the oxidation of reduced sedimentary rocks. The sink side corresponds to the storage of carbon on the seafloor either by carbonate sedimentation (F_{cd}) or by organic carbon burial into the sediment (F_{od}).

(2) The alkalinity balance:

$$F_{swo} + F_{swy} + F_{cw} = F_{cd} \tag{A2}$$

F_{swo} and F_{swy} are respectively half of the alkalinity flux created during the weathering of continental old cratonic silicate rocks and of young volcanic rocks. These two terms can be written as follows:

$$F_{swo} = \left(1-\sigma_y\right)F_{sw} \tag{A3}$$

$$F_{swy} = \sigma_y F_{sw} \tag{A4}$$

where F_{sw} equals half of the total alkalinity flux created during the weathering of continental silicate rocks. σ_y is the fraction of the silicate rock outcrop occupied by young volcanic rocks.

Altogether, these two Eqs. (A1) and (A2) can be written as follows, stating that all sources of exospheric carbon by solid Earth degassing is compensated by the continental silicate weathering and the net sink of organic carbon (organic carbon burial minus the oxidation of reduced sedimentary carbon exposed at the Earth surface).

$$F_{vol} + F_{MOR} = F_{swo} + F_{swy} + (F_{od} - F_{ow}) \tag{A5}$$

(3) The organic carbon disequilibrium ($F_{od} - F_{ow}$) can be constrained by considering the carbon isotopic balance.

$$\begin{aligned}\delta_{vol}F_{vol} + \delta_{MOR}F_{MOR} + \delta_{CW}F_{CW} + \delta_{OW}F_{OW} \\ = \delta_{oc}F_{cd} + \left(\delta_{oc}-\varepsilon_p\right)F_{od}\end{aligned} \tag{A6}$$

where δ_{vol} is the averaged volcanic gas $\delta^{13}C$ signature, and δ_{MOR} is the mantle carbon isotopic composition. δ_{cw} and δ_{ow} are the averaged $\delta^{13}C$ of continental carbonated and reduced sedimentary rocks. δ_{oc} is the measured seawater $\delta^{13}C$ (Veizer et al., 1999) and ε_P is the photosynthetic carbon isotopic fractionation (Hayes et al., 1999).

(4) The strontium isotopic budget can be used to constrain the relative importance of young volcanic rocks versus older silicate rocks (Berner, 2004). Given the residence time of strontium in seawater (1.5 to 2 Ma) and the time resolution

of the model simulations (at least several tens of million years), we assume this budget to be close to steady state:

$$\frac{r_{MOR}-r_{OC}}{9.43+r_{MOR}}F_{MOR}^{Sr}+\frac{r_{CW}-r_{OC}}{9.43+r_{CW}}F_{cw}^{Sr}+\frac{r_{swy}-r_{OC}}{9.43+r_{swy}}F_{swy}^{Sr}+\frac{r_{swo}-r_{OC}}{9.43+r_{swo}}F_{swo}^{Sr}=0 \quad (A7)$$

where r_{MOR}, r_{cw}, r_{swy}, and r_{swo} are the $^{87}Sr/^{86}Sr$ ratio of the mantle, carbonates, young volcanics and old cratonic rocks. These are set to 0.703, 0.708, 0.703 and 0.718 respectively. r_{OC} is the measured $^{87}Sr/^{86}Sr$ ratio of seawater recorded in sedimentary carbonates (Veizer et al., 1999). The superscript *Sr* to the flux *F* means that we are considering the strontium fluxes associated to the carbon and alkalinity fluxes.

(5) Lastly, three additional equations describe the dependence of silicate (F_{swy} and F_{swo}) and carbonate weathering (F_{cw}) on atmospheric CO_2, through their dependencies on runoff and air temperature. In GEOCARB-style models, only global runoff and mean average global temperature are calculated as a function of CO_2 (Eqs. (1) and (2)), and then used to calculate mean global weathering fluxes. In GEOCLIM, a physically-based 3D climate model is used to calculate the spatial distribution of runoff and temperature as a function of CO_2, which are subsequently used to reconstruct spatially-distributed weathering fluxes.

If the solid Earth degassing rate is known (F_{vol} and F_{MOR}), the 9 Eqs. ((A1), (A2), (A4), (A5), (A6), (A7) + the three climatic dependences) can be solved for F_{cw}, F_{ow}, F_{sw}, F_{swy}, F_{swo}, σ_y, F_{cd}, F_{od}, and atmospheric CO_2 (9 unknowns).

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