The role of shelf nutrients on glacial-interglacial CO2: A negative feedback
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[1] In the past 800 thousand years and before industrialization, the largest variations in atmospheric CO2 concentration (pCO2) occurred in connection with the glacial cycles that characterized Earth’s climate over this period. One curious feature of at least the last four glacial-interglacial cycles is that atmospheric pCO2 reached about the same upper limit of 280 ppm during peak interglacial periods and about the same lower limit of 180 ppm during peak glacial periods. Here, we show using a numerical model of earth system that enhanced shelf sediment weathering during glacial sea level lowstands tends to raise pCO2 even after carbonate compensation and thus stabilize pCO2 from further reduction. This is because not all nutrients from weathering will be utilized by biology but more importantly because the spatial distributions of carbon and phosphorus from weathering become decoupled in such a way that carbon is preferentially stored in the upper ocean and phosphorus in the deep ocean. In addition, the C:P ratios in continental margin sediments are generally much higher than the Redfield ratio due to preferential remineralization of phosphorus in shelf sediment diagenesis. When these factors are accounted for in our model, the input of organic matter, which corresponds to the observed negative shift in ocean δ13C during glacial periods, raises pCO2 by approximately 14 ppm. The same mechanisms operating in the opposite directions during interglacial highstand tend to lower pCO2 and stabilize it from further increase. The impact of sea level-driven continental shelf exposure and submersion of CO2 is therefore a negative feedback that may have contributed to limiting the variation of Pleistocene pCO2 to the observed 100 ppm range.


1. Introduction

[2] Atmospheric pCO2 is one of the key factors regulating the global climate, and the relationship between pCO2 and climate has been of great interest in many researches. The importance of CO2 on climate as one of the most influential greenhouse gases is clear, but there is as yet no widely accepted explanation for how pCO2 could have varied by as much as 100 ppm over the glacial-interglacial cycles [e.g., Archer et al., 2000; Sigman and Boyle, 2000; Toggweiler et al., 2006; Sigman et al., 2010]. In addition, the reason why this pCO2 variation was largely restricted to the range between 180 and 280 ppm over the last four glacial cycles is a mystery. The pCO2 variation in earlier glacial cycles was apparently somewhat smaller as was the air temperature variation over Antarctica [Siegenthaler et al., 2005].

[3] In the so-called shelf nutrient hypothesis, Broecker [1982] proposed that the exposure of continental shelf sediments during glacial sea level lowstands would have increased the weathering input of nutrients to the ocean, enhanced its biological production, and subsequently caused a drawdown of CO2 from the atmosphere. This hypothesis, as well as the coral reef hypothesis [Berger, 1982], both fell out of favor rather quickly largely because they require the trigger, the global sea level, to change before pCO2, but this relative timing is not supported by available observations [Broecker and Henderson, 1998]. However, these hypotheses could still play a role in the glacial pCO2 mystery, because the field now recognizes that there were likely multiple mechanisms of glacial-interglacial pCO2 variability [Archer, 2004] thus permitting drivers of pCO2 variation to have acted asynchronously. In fact, it seems difficult to imagine that the enormity of the event, a 120 m change in eustatic sea level between peak glacial and interglacial times, did not have any impact on atmospheric pCO2.

[4] In order to quantify the impact of shelf sediment exposure and weathering on pCO2, we conduct a series of numerical experiments using an earth system model of intermediate
complexity. We focus on two aspects of the shelf nutrient hypothesis: the amount and C:P ratio of the organic matter weathered. The amount reflects the magnitude of the perturbation. The importance of the C:P ratio is easily understood when considering extreme cases. For example, if only phosphorus (C:P = 0) entered the ocean, any new production based on the excess phosphorus will lead to reduction in pCO₂. On the other hand, if the weathered organic matter had a C:P ratio higher than the mean stoichiometry of organic matter produced in the global ocean, even a complete utilization of the excess phosphorus would leave behind excess carbon, thus raising atmospheric pCO₂. Our finding that the impact on pCO₂ is opposite of what was originally envisioned by the shelf nutrient hypothesis has a potentially significant implication for glacial-interglacial pCO₂ variation and amplitude.

2. Method (Design of Experiments)

2.1. Model Description

[5] We use the Minnesota Earth System Model for Ocean biogeochemistry (MESMO 1.0), a 3-D intermediate complexity global model of biogeochemistry and climate [Matsumoto et al., 2008]. It is derived from the climate model of Edwards and Marsh [2005] and has three closely integrated modules: a 3-D ocean circulation model, a dynamic thermodynamic model of sea ice, and a 2-D energy-moisture balance model of the atmosphere. The dynamics of the ocean model is based on the frictional geostrophic equations as described by Edwards and Marsh [2005]. It has a 36 × 36 equal area horizontal grid with 10° increments in longitude and is uniform in size of latitude; latitude spacing increases from about 3° at the equator to about 20° at the poles. There are 16 levels in the vertical with the top two layers being 100 m thick. The biogeochemical module of MESMO includes a simple prognostic export production that depends on Michaelis-Menten nutrient uptake kinetics, light, estimated phytoplankton biomass, temperature, and mixed layer depth. Remineralization of particles is based on a tuned rate of sinking and temperature-dependent rate of remineralization. Also, seasonal variation in insolation causes seasonal variation in export production, as light availability, mixed layer depth, and vertical nutrient supply change. Key biogeochemical parameters have been tuned using oxygen and global production as targets.

[6] Export production of carbon, nitrogen, and phosphorus is related by elemental stoichiometry of C:N:P = 117:16:1 according to Anderson and Sarmiento [1994]. The production of CaCO₃ occurs in waters supersaturated with carbonate ion with respect to mineral calcite. It is related to organic carbon production by a ratio, which depends on the degree of supersaturation, so that in lower pH waters, CaCO₃ production is reduced and thus organic carbon to CaCO₃ export ratio becomes higher [Ridgwell et al., 2007].

[7] A 1-D sediment geochemistry model (SEDGEM) is coupled at each grid of ocean floor to represent the accumulations of CaCO₃,opal and clay [Ridgwell and Hargreaves, 2007; Chikamoto et al. 2008]. It has a 5 cm-single surface layer underlain by a stack of sediment storage layers. Each sediment module is driven by particle settling fluxes as well as ambient temperature and salinity, and the carbonate chemistry in the overlying ocean. The dissolution flux of CaCO₃ is obtained through linear interpolation on a regular multidimensional grid of precalculated values [Archer, 1991]. Oceanic inventories of alkalinity and dissolved inorganic carbon (DIC) are calculated explicitly by the balance between inflow of riverine CaCO₃ flux (10 Tmol yr⁻¹, based on the estimation of Ridgwell and Hargreaves [2007]) and outflow of sediment burial flux as calculated by the sediment model. Organic matter is assumed to be entirely remineralized in the surface sediments of the deep sea and so omit organic matter as an explicit component of accumulating sediments. The introduction of the sediment model and river inputs allows the model to simulate open system responses including carbonate compensation.

[8] The control run uses the boundary conditions of pre-industrial era and has pCO₂ of 278 ppm. It is well calibrated in terms of biological production and with respect to standard metrics of the modern ocean ventilation, including anthropogenic carbon, chlorofluorocarbon, and natural radiocarbon [Matsumoto et al., 2008]. For example, the oceanic uptake of anthropogenic carbon and CFC-11 for the year 1994 is respectively 120 Pg-C and 0.68 × 10⁹ moles, which compare favorably to data-based estimates of 118 ± 19 Pg-C and 0.55 ± 0.12 × 10⁹ moles. These transient tracer uptake would compare well against those in the Ocean Carbon Cycle Model Intercomparison Project [Matsumoto et al., 2004]. The model is successfully used in a number of process studies of the global carbon cycle of the present and glacial periods [Chikamoto et al., 2008; Matsumoto et al., 2010; Sun and Matsumoto, 2010; Lee et al., 2011] as well as in community-wide model intercomparison projects [Archer et al., 2009; Cao et al., 2009].

2.2. Experiment Design

[9] The effect of weathering of organic matter from exposed continental shelves is simulated in MESMO by introducing dissolved organic matter (DOM) to the ocean model domain along the coast, where runoff is expected from the model’s land surface runoff scheme [Matsumoto et al., 2010]. DOM is composed of dissolved organic carbon (DOC), nitrogen (DON), and phosphorus (DOP). In a series of experiments, we explore a parameter space where the amount of DOM entering the ocean and its C:P ratio are varied. In all sensitivity experiments, DOM is introduced in the first 200 years. After the system reaches equilibrium, the model state is compared against control run, in which no DOM was introduced. The equilibrium state is independent of the rate and duration of the DOM input, so long as the input does not occur too rapidly and causes the carbonate chemistry calculation to become numerically unstable.

[10] The magnitude of perturbation, or the amount of DOM input, is expressed in terms of carbon and prescribed to one of three values: 10, 300, or 600 Pg-C. Glacial-interglacial change in the whole ocean δ¹³C provides a constraint on how much organic carbon could have been added to the ocean. Following the seminal work by Shackleton [1977], a number of subsequent studies have quantified the δ¹³C change [e.g., Curry et al., 1988; Duplessy et al., 1988; Crowley, 1995]. Bird et al. [1994] estimate that 310–550 Pg of organic carbon input is required to satisfy the observed 0.32‰ δ¹³C change by a simple isotope mass balance calculation. Although larger estimates on terrestrial carbon storage variation have been made from terrestrial biome reconstruction [e.g., Adams et al., 1990], most estimates of glacial carbon storage change
converge in a range between 300 and 700 Pg-C [e.g., Friedlingstein et al., 1992; Prentice et al., 1993; Van Campo et al., 1993; Friedlingstein et al., 1995; Street-Perrott et al., 1997; Beerling, 1999; Ikeda and Tajika, 2003]. These values are usually considered to reflect land-derived organic carbon, not separated from shelf organic carbon. In this study, a 600 Pg input of organic carbon ($\Delta^{13}C = -25\%$) serves as the upper limit of organic carbon input to the ocean as it causes a shift of ocean $\Delta^{13}C$ by $-0.3\%$ in our model.

The C:P ratio of DOM entering the ocean model domain is varied between 0 and 480, which encompasses the modern ratio of 117, while the N:P ratio is fixed to 16 in all experiments. Elemental composition of organic matter in shallow sediments is poorly constrained by observations, but available studies indicate that phosphorus is preferentially remineralized, so that C:P of remaining organic matter in sediments generally becomes higher with increasing diagenesis. Hartmann et al. [1976] observed C:P ratios in sediment organic matter increased with increasing depth in rapidly depositing sediments. Filipke and Owen [1981] and Krom and Berner [1981] also reported C:P ratios greater than 200 in shallow (9, 30 and 112 m) sediments. For simplicity, we first illustrate the behavior of carbon and phosphorus released to the ocean with a ratio of 117, which is the C:P stoichiometry for new production in MESMO, before considering more realistic cases of higher C:P ratios for continental margin sediments.

3. Results

It seems at first glance that there would not be any change in atmospheric $pCO_2$, if phytoplankton fixed phosphorus and carbon in the same ratio as that in weathering. Contrary to this expectation however, we find that any input of DOM with the same C:P ratio of 117 as the ratio in new production results in atmospheric $pCO_2$ rise (Figure 1). The rapid and initial increase is a result of $CO_2$ escaping the surface ocean, which becomes quickly supersaturated with $CO_2$ as DOM from weathering decomposes into DIC. Subsequently, atmospheric $CO_2$ declines as ocean circulation redistributes the perturbation carbon throughout the ocean, reducing surface ocean $pCO_2$ and facilitating uptake of the perturbation $CO_2$ that escaped to the atmosphere. Meanwhile, biological production also modifies the distributions of phosphorus, carbon, and alkalinity. On longer time scale and in this case, sedimentary carbonate compensation tends to lower atmospheric $pCO_2$ as the ocean responds to DOM input, which basically acidifies the ocean, by dissolving sedimentary CaCO$_3$ and restoring deep ocean carbonate ion concentration. When a new equilibrium is established after 20,000 years, $pCO_2$ remains above the initial value of 278 ppm. In fact, $pCO_2$ increases even when the C:P ratio of DOM input is well below 117 as discussed below.

In the following sections, we first explore the reasons for this unexpected response with MESMO without sediment geochemistry, since carbonate compensation modifies but does not change the sense of the model response. It is easier to elucidate the main reasons without the sediments, whose reactions are not easily controlled in a set of sensitivity runs we discuss below.

3.1. Water Column Response and Its Mechanisms

Experiment runs without sediments reach new equilibriums after 3,000 years. Figure 2 shows the new states as compared to the control run within a parameter space defined by DOM inputs ranging from 0 to 600 Pg-C and by C:P ratios of those inputs ranging from 0 to 500. Atmospheric $pCO_2$ increases as both DOM input and C:P ratio increase (Figure 2a). However, following the result shown in Figure 1, it is not until the C:P ratio is below 60 that $pCO_2$ decreases. Surface PO$_4$ increases in proportion to the amount of introduced DOP (Figure 2b). Surface alkalinity is largely a function of CaCO$_3$ production, which in turn depends on surface pH. Alkalinity at the surface tends to

Figure 1. Model $pCO_2$ responses to DOM inputs. Model results show rapid elevation of $pCO_2$ over the first 200 years in response to sudden input of organic matter with different C/P ratios and total C contents into surface coastal region, before reaching a new equilibrium state as added carbon and nutrients are consumed by production and diffused globally by the ocean circulation. After 20,000 years, the system reaches new equilibrium states and shows elevated $pCO_2$ even by input of DOM with the elemental stoichiometry of C:N:P = 117:16:1.
increase as the surface ocean becomes more acidic (lower pH) in response to greater DOM input and higher C:P ratio (Figure 2c). Surface DIC, which is determined primarily by organic carbon production and air-sea gas exchange, also generally increases with greater DOM input and C:P ratio (Figure 2d).

[15] In order to grasp why atmospheric CO₂ increases even when C:P of DOM entering the ocean is the same as the ratio of new production, we separately quantified the abiotic effects from biotic effects on atmospheric CO₂ in a series of runs without sediment geochemistry, where organic carbon and CaCO₃ productions were allowed to respond in stepwise manner (Figure 3a). We analyze the sensitivity run with DOM input of 600 Pg-C and C:P of 117, which resulted in final atmospheric pCO₂ of 301 ppm, a 23 ppm increase from the initial state (Figures 2a and 3a).

[16] In the first phase of the experiment, both organic carbon and CaCO₃ productions are maintained at the control run levels as weathering inputs occur. This was accomplished by applying monthly production masks, diagnosed from the initial equilibrium state. In this constant biology phase, pCO₂ quickly increases by 153 ppm to 431 ppm and then declines slowly to reach 336 ppm in 3,000 years (Figure 3a, left). These changes are driven abiotically, first as the surface ocean becomes supersaturated with CO₂ and later as ocean circulation redistributes DIC throughout the

Figure 2. Contour plots of responses of (a) pCO₂, (b) surface PO₄, (c) surface alkalinity, and (d) surface DIC to input of organic matter with different C/P ratios and total C contents into surface coastal region in MESMO without sediment geochemistry. Shown are anomalies from the control run after 3,000-year integration. Red and blue gradations indicate positive and negative directions of changes, respectively. Black dots indicate conditions of C/P ratios and total C contents that perturbation experiments are conducted with.
ocean. If all phosphorus and carbon from weathering remained in the ocean, surface ocean PO$_4$ and DIC would change in proportion to the C:P ratio in the weathered DOM. This hypothetical case is indicated by point A' and dashed line with slope of 117 in Figure 3b. However, since some carbon is degassed from the ocean and does not return, thus raising atmospheric pCO$_2$, the ratio of change in carbon to phosphorus from the initial equilibrium state ($D_{DIC}/D_{PO4}$) in surface waters is lower than 117 (Point A in Figure 3b). Open circles (points A' and B') indicate hypothetical cases where surface DIC change occurs in proportion to PO$_4$ change, as indicated by the dashed line with slope of 117. With sediment geochemistry and carbonate compensation, the final pCO$_2$ would likely be 282 ppm (above the initial 278 ppm).

In the second phase, we allow organic carbon production to respond to weathering inputs. The extra phosphorus fuels greater production, which reduces atmospheric pCO$_2$ by 38 ppm from points A to B (Figure 3a, middle). Organic carbon production draws down surface DIC and phosphorus in the prescribed stoichiometry of 117 along the dashed line to point B' in Figure 3c. The reduction in surface ocean DIC in turn drives air-to-ocean flux of CO$_2$, so that surface $\Delta$DIC/$\Delta$PO$_4$ ends up becoming higher than the input C:P ratio of 117 (Point B in Figure 3c). This decoupled behavior of carbon and phosphorus is the main cause of why atmospheric pCO$_2$ becomes elevated even when the input DOM has a C:P ratio that is identical to the ratio of new production. The other reason is that not all phosphorus from weathering is fully utilized by biology, which is evident from the fact that surface PO$_4$ concentration after organic carbon production is higher than the initial concentration (Figures 2b and 3c).

In the third and final phase, we allow CaCO$_3$ production to respond, which increases atmospheric pCO$_2$ since CaCO$_3$ production removes twice as much alkalinity as DIC thus reducing the carbonate ion concentration and increasing pCO$_2$. Therefore atmospheric pCO$_2$ increases by 3 ppm (Figure 3a, right) and DIC decreases from point B to point C in Figure 3d. The final pCO$_2$ increase is 23 ppm above the initial 278 ppm.

3.2. Effects of Carbonate Compensation

Unfortunately, these stepwise sensitivity experiments were not possible in a model with sediment geochemistry, as the sedimentary feedbacks cannot be controlled and the
control runs for MESMO with and without the sediments are not the same. However, we are able to obtain the final, steady state solutions with sediment geochemistry, sweeping the same parameter space as in Figure 2. Figure 4 shows the results after 20,000 years of model integration that explicitly accounts for carbonate compensation, which in this case drives deep-sea CaCO₃ dissolution and tends to restore deep ocean carbonate ion concentration. Compared to Figure 2, Figure 4 shows that carbonate compensation reduces atmospheric pCO₂, increases surface alkalinity and DIC, and leaves surface PO₄ concentration unchanged. For the equivalent experiment shown in Figure 3 (600 Pg-C input and with C:P ratio of 117), carbonate compensation diminishes the final atmospheric pCO₂ from 23 ppm above the initial 278 ppm to 4 ppm above 278 ppm. Also, the C:P ratio at which the change in atmospheric pCO₂ reverses sign is elevated from 60 to 90 (see Figure 4a versus Figure 2a). The overall effect of carbonate compensation on CO₂ is therefore to reduce but not erase the water column effects identified in section 3.1.
3.3. Decoupled Behavior of Carbon and Phosphorus

The decoupled behavior of the added carbon and phosphorus is indicated by $\Delta \text{DIC}/\Delta \text{PO}_4$, and its spatial variability also sheds light on why atmospheric $p\text{CO}_2$ increases even when the input C:P is 117 (Figure 5). The drawdown of carbon and phosphorus by organic carbon production combined with invasion of CO$_2$ from the atmosphere, illustrated in Figure 3c, causes the mean surface ocean $\Delta \text{DIC}/\Delta \text{PO}_4$ to increase to almost 230 (Figure 5a). However, $\Delta \text{DIC}/\Delta \text{PO}_4$ rapidly decreases with depth to values lower than 117 within the top 300 m (Figure 5a). This means that perturbation carbon and phosphorus are vertically fractionated, such that carbon is preferentially stored in the upper ocean and phosphorus in the deep. The low $\Delta \text{DIC}/\Delta \text{PO}_4$ in the deep ocean reflects the fact that $\Delta \text{DIC}/\Delta \text{PO}_4$ at the surface is low in high latitudes where deep water forms (Figures 5b and 5c). The surface variability is a result of smaller biological uptake of phosphorus and greater Revelle buffer factor in higher latitudes. The higher Revelle factor in the high latitudes means that DIC changes less for a given change in $p\text{CO}_2$ than the lower Revelle factors in low latitudes [Takahashi et al., 1980]. These two primary factors keep surface $\Delta \text{DIC}/\Delta \text{PO}_4$ smaller in higher latitudes than in lower latitudes and consequently maintain a large vertical gradient in $\Delta \text{DIC}/\Delta \text{PO}_4$.

4. Discussion

In this study, the input of DOM into the ocean results in preferential accumulation of carbon relative to phosphorus in the surface ocean, as indicated by the spatial variability of $\Delta \text{DIC}/\Delta \text{PO}_4$ (Figure 5). This is driven by the decoupling of carbon through air-sea gas exchange from phosphorus (Figure 3). The preferential accumulation of carbon at the surface, combined with incomplete biological consumption of phosphorus input from weathering, causes CO$_2$ level to increase in the atmosphere even when the weathering input has C:P ratios well below that of new production (Figure 2). Carbonate compensation tends to counter this increase in atmospheric $p\text{CO}_2$ but is not large enough to erase it (Figure 4).

[21] It is interesting to note that the original shelf hypothesis by Broecker [1982] was apparently able to reduce atmospheric $p\text{CO}_2$, while assuming the same C:P ratio in the shelf sediments as in new production. For one, it was
assumed that input nutrients were entirely consumed, which is shown in this study to be unrealistic (Figures 2b and 4b). More importantly however, Broecker [1982] assumed that the lysocline depth did not change and explicitly introduced the same amount of CaCO3 as organic carbon so as to maintain deep ocean carbonate ion concentration. In a tabular calculation, this is a way to represent carbonate compensation and increased ocean alkalinity (and DIC) in such a way that atmospheric pCO2 is ultimately reduced when all excess phosphorus is consumed by phytoplankton. In his calculation, the lower glacial pCO2 was actually a result of carbonate compensation rather than the input of organic matter.

[23] In contrast to the original shelf nutrient hypothesis, this study indicates that DOM input from continental shelf weathering during glacial periods would in fact raise atmospheric pCO2, because C:P ratio in shelf sediments is typically higher than in new production as remineralization preferentially removes phosphorus. Carbonate compensation cancels a sizable part of pCO2 increase but not entirely. We therefore argue that the shelf nutrient weathering does not help explain why pCO2 became low during glacial periods but acts as a negative feedback on pCO2, preventing it from falling further. It is difficult to pinpoint the strength of this negative feedback at specific times in the past, because the C:P ratio of shelf sediments and the amount of DOM weathered during sea level lowstands are poorly constrained. However, available literature indicate that C:P ratio in shallow sediments are generally much higher than the Redfield ratio [Ruttenberg and Berner, 1993], and the ratio can be higher than 800 in sediments overlain by oxygen-depleted waters [Ingall and Jahnke, 1997]. If we assume that the C:P ratio was as high as 200 and that 50–100% of the organic matter needed to account for the 0.3–0.4‰ shift in benthic foraminiferal δ18O [Duplessy et al., 1988; Crowley, 1995] came from the shelves, our results would indicate that atmospheric pCO2 would increase by 5–14 ppm.

[24] When could this sea level-driven negative feedback have taken effect? In order to speculate on this, we calculated the area of continental shelf that would have been exposed as sea level fluctuated over the past 400 kyr using benthic foraminiferal δ18O stack data [Lisiecki and Raymo, 2005] as a sea level proxy and accounting for hypsometry using global elevation data ETOPO1 [Amante and Eakins, 2009]. Comparing it with pCO2 reconstructed from the Vostok ice core, we suggest that perhaps the rate of pCO2 reduction slowed when the exposed continental shelf area becomes very large (e.g., shading in Figure 6 indicates exposure greater than 70% of global continental shelf within 150 m depth of the present sea level). This typically occurs in the latter half of the 100 kyr march toward glacial maximum and termination, so that sea level and atmospheric pCO2 have already dropped substantially by this time and the lack of large variability thereafter is consistent with our speculation that shelf sediment weathering during glacial sea level lowstands dampened a further reduction in pCO2.

[25] There is a corollary to this speculation that operates during the interglacial period (Figure 7). During interglacial highstands, the preferential recycling of phosphorus relative to carbon in shelf sediments not only gives them high C:P ratios but also acts as an effective carbon sink. Phosphorus can shuttle back and forth between the sediment and water column, taking carbon on its way down but returning up into

**Figure 6.** Time series of the area of exposed continental shelf (black solid line, area of 0–150 m depth) and pCO2 (gray dotted line) over the last 420,000 years. The exposed area of continental shelf was calculated from benthic foraminiferal δ18O stack data of Lisiecki and Raymo [2005] and pCO2 was reconstructed from the Vostok ice core [Petit et al., 1999]. Shaded areas represent high percentage (>70%) of shelf exposure. We note that the estimation of sea level using benthic δ18O during earlier interglacial highstands is likely not accurate as deep water was likely warmer [e.g., Kutler et al., 2003]; however, such periods would be outside the shaded areas of interest. Therefore, the inaccuracy would be largely irrelevant to this study.

**Figure 7.** Schematic diagrams of the role of continental shelves on pCO2 during peak interglacial and peak glacial periods. (a) During peak interglacial, phosphorus is remineralized preferentially and released back from shelf sediments to the water column, acting as a CO2 sink. (b) During peak glacial, as shelf sediments generally have higher C:P ratio, an enhanced weathering of such sediments makes the ocean release CO2, as shown in this study.
the water column carbon-free. This can also be viewed as a negative feedback, which prevents interglacial CO$_2$ from rising further.

5. Conclusions

[26] We have identified a possible negative feedback on atmospheric CO$_2$ that helps maintain its variability to the roughly 100 ppm that is observed in ice cores. When weathered carbon and phosphorus are added to the ocean-atmosphere system, phosphorus will stay in the ocean whereas the carbon will be redistributed between both the ocean and atmosphere. Our model simulations suggest that atmospheric CO$_2$ would increase when continental shelves are exposed and weathered during glacial lowstands; this suggestion is opposite of the original shelf nutrient hypothesis of Broecker [1982].

Our explicit simulation of carbonate compensation indicates that it was not as dominant as assumed originally in a tabular calculation. An implication of this study is that during interglacial highstands, phosphorus may effectively sequester carbon in the shelf sediments as it is preferentially remineralized during diagenesis and becomes repeatedly available for biological carbon pump. The tendency for shelf exposure to increase CO$_2$ during glacial periods and shelf submersion to reduce it during interglacial periods would qualify it as a negative feedback.

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