The impact of remineralization depth on the air–sea carbon balance

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As particulate organic carbon rains down from the surface ocean it is respired back to carbon dioxide and released into the ocean’s interior. The depth at which this sinking carbon is converted back to carbon dioxide—known as the remineralization depth—depends on the balance between particle sinking speeds and their rate of decay. A host of climate-sensitive factors can affect this balance, including temperature, oxygen concentration, stratification, community composition and the mineral content of the sinking particles. Here we use a three-dimensional global ocean biogeochemistry model to show that a modest change in remineralization depth can have a substantial impact on atmospheric carbon dioxide concentrations. For example, when the depth at which 63% of sinking carbon is respired increases by 24 m globally, atmospheric carbon dioxide concentrations fall by 10–27 ppm. This reduction in atmospheric carbon dioxide concentration results from the redistribution of remineralized carbon from intermediate waters to bottom waters. As a consequence of the reduced concentration of respired carbon in upper ocean waters, atmospheric carbon dioxide is preferentially stored in newly formed North Atlantic Deep Water. We suggest that atmospheric carbon dioxide concentrations are highly sensitive to the potential changes in remineralization depth that may be caused by climate change.

The downward flux of particulate organic carbon, $F$, at a depth $z$ below the base of the euphotic zone $z_i$ is often represented by the power law function $F(z) = F(z_i) \times (z/z_i)^{-b}$ (ref. 6) where the exponent $b$ controls the efficiency of transfer of particulate organic carbon to depth. The global mean value of the exponent $b$ estimated using ocean biogeochemical tracer data is 0.9–1.0 (refs 7–9). Sediment trap observations indicate that the downward transfer efficiency may vary regionally and temporally with the exponent $b$ ranging from 0.5 to 2.0 (refs 4, 11–14). Previous studies using three-dimensional global ocean biogeochemistry models showed that the remineralization depth has an important role in controlling the global distributions of nutrients and carbon. In ref. 7, a large change in atmospheric $pCO_2$ (partial pressure of carbon dioxide) of $\sim 100$ ppm was obtained when the exponent $b$ was changed from 0.9 to 2.0. We show here that even a small change in remineralization depth results in a substantial change in atmospheric $pCO_2$ and we explore the underlying mechanisms of this large sensitivity.

We performed a systematic sensitivity analysis of the distributions of nutrients and carbon to changes in remineralization depth using a 3D global ocean biogeochemistry model (see the Methods section) coupled with a one-box model of the atmosphere. Because the response of the new production of organic matter to nutrient supply is uncertain, we bracketed the ocean’s response to changes in the remineralization profile by considering two extreme cases: a ‘nutrient-restoring’ model where new production depends entirely on the supply of nutrients to the surface ocean by advection and mixing and a ‘constant-export’ model where new production is kept fixed unless the $PO_4$ concentration in the production zone drops to zero (see the Methods section). The latter model is more appropriate if new production is limited by external factors such as light or micronutrient input from wind-blown dust.

The effect of changing remineralization depth on the distribution of nutrients is illustrated in Fig. 1a. With increase in the remineralization depth, respired nutrients are delivered deeper in the water column, leading to a net transfer of $PO_4$ from the upper ocean to the deeper ocean. The largest $PO_4$ increase is in the bottom waters of the North Pacific Ocean whereas North Atlantic Deep Water (NADW) shows a pronounced decrease in $PO_4$. The maximum $PO_4$ increase in the deep North Pacific Ocean reflects the increased accumulation of respired nutrients in the deep branch of the global overturning circulation. The reduction of $PO_4$ in newly formed NADW can also be understood in terms of the interaction of the global overturning circulation and the shift of respired nutrients from intermediate to bottom waters. The upper branch of the Great Ocean Conveyor carries intermediate waters with reduced nutrient concentrations into the North Atlantic Ocean where it is then exported southward as NADW (Fig. 1b). NADW therefore has the effect of decreasing the inventory of preformed $PO_4$ (biologically unused $PO_4$ transported from the surface) in the deep ocean (Supplementary Fig. S1).

The downward shift of respired nutrients results in an increased global mean concentration of remineralized $PO_4$, although export production of organic material either decreases or does not change (Fig. 2a,b). The pool of remineralized $PO_4$ increases because the mean residence time of remineralized nutrients in the ocean’s interior increases. Assuming a constant stoichiometric ratio of C:P, the increase in remineralized $PO_4$ implies an increase in the pool of respired carbon, which, in turn, implies a decrease in atmospheric $pCO_2$ (refs 19, 20). Our simulations show substantial variations in atmospheric $pCO_2$ associated with the rearrangement of the $PO_4$ distribution (Fig. 2c). For example, $pCO_2$ decreases by $86$ ppm in the nutrient-restoring model and by $185$ ppm in the constant-export model as the exponent $b$ changes from 1.4 to 0.5.

We can associate the exponent $b$ with the remineralization e-folding depth, the depth by which 63% of organic matter exported from the euphotic layer has become remineralized. Figure 2d shows that the sensitivity of atmospheric $pCO_2$ is strongest for e-folding depth scales of less than a few hundred metres. Because the present best estimate for the exponent $b$, which corresponds to an e-folding depth of $210$ m, lies well within the high sensitivity range, a small perturbation of the remineralization depth from its present value can lead to a considerable change in the air–sea carbon partitioning. For example a change of the e-folding depth from 204 to 228 m...
in the nutrient-restoring model leads to a decrease in atmospheric pCO$_2$ of 10 ppm and the same change in the constant-export model leads to a decrease of 27 ppm.

To elucidate the mechanisms by which a deepening of the remineralization depth leads to carbon uptake, we divide the total response of dissolved inorganic carbon ($\Delta$DIC) into contributions from the soft-tissue pump ($\Delta$DIC$_{soft}$), the carbonate pump ($\Delta$DIC$_{carb}$) and the gas-exchange pump ($\Delta$DIC$_{gas}$) (refs 8, 21; see the Methods section). $\Delta$DIC$_{soft}$ and $\Delta$DIC$_{carb}$ represent the changes in DIC that would occur if there were only organic carbon and calcium carbonate (CaCO$_3$) fluxes, respectively, in the absence of air–sea gas exchange. Because the soft-tissue and carbonate pumps (as defined in ref. 21) only rearrange DIC within the ocean, the global integrals of $\Delta$DIC$_{soft}$ and $\Delta$DIC$_{carb}$ are zero. The remaining $\Delta$DIC$_{gas}$ therefore reveals the distribution of the excess atmospheric CO$_2$ taken up by the ocean. We also divide the total response of alkalinity into contributions from the soft-tissue and carbonate pumps$^8$ (see the Methods section) to help understand the causes of carbon uptake.

Figures 3 and 4 illustrate how an increase in the remineralization depth affects the air–sea carbon balance through changes in the soft-tissue and carbonate pumps. An increase in the remineralization depth results in redistributing $\Delta$DIC$_{soft}$ from relatively well-ventilated waters to ones that are poorly ventilated. At the same time, surface alkalinity increases due to a reduction in remineralized nitrate in the upper ocean. The decrease in surface DIC and the increase in surface alkalinity both cause the ocean to take up CO$_2$ from the atmosphere.

A weakening of the carbonate pump also contributes to carbon uptake when the remineralization depth increases. With increased remineralization depth, the amount of CaCO$_3$ exported from the production zone decreases globally at the same rate as the reduction of particulate organic carbon export (Fig. 2b and Methods). The reduced production of CaCO$_3$ results in increased carbonate ion...
concentrations in the upper ocean and decreased concentrations in the deeper ocean. This rearrangement causes the ocean to take up CO2 due to the increase in surface alkalinity.

The mechanism mediated by the soft-tissue pump accounts for ~85% of the pCO2 drawdown in the constant-export model and ~50% in the nutrient-restoring model (Fig. 2c,d). With the CaCO3 cycle turned off, for example, an increase in the remineralization e-folding depth from 204 to 228 m leads to decreases of atmospheric pCO2 by 23 ppm for the constant-export model and 5 ppm for the nutrient-restoring model. The CO2 uptake mediated by the soft-tissue pump is about five times greater in the constant-export model, where new production decreases minimally in response to a reduction in surface nutrients.

The global ocean increases its capacity for CO2 when the remineralization depth increases. Figure 3 shows that increases in ΔDICcos are particularly pronounced in the water formed in the high-latitude North Atlantic Ocean. The increased accumulation of respired nutrients in old bottom waters is balanced by the depletion of preformed nutrients in newly formed NADW, which causes carbon accumulation taken in NADW. A net transfer of carbonate ions from old bottom waters to NADW also contributes to CO2 uptake in the NADW formation regions. As a result, NADW provides a major pathway of atmospheric CO2 to the deep ocean when the remineralization depth increases globally (Fig. 3b; see also Supplementary Information).

The Southern Ocean has been recognized as a region within which the efficiency of the biological pump controls atmospheric pCO2. Owing to inefficient biological consumption of surface nutrients, water masses ventilated from the Southern Ocean are characterized by high concentrations of preformed nutrients (average 1.6 µmol kg⁻¹ of PO4, compared to 0.9 µmol kg⁻¹ in NADW)¹⁹. Thus an increased biological use of surface nutrients leads to net uptake of carbon primarily through bottom waters formed in the Southern Ocean²². However, increasing remineralization depth does not deplete surface nutrients in the Southern Ocean as much as it does in the NADW formation regions. This is because Circumpolar Deep Water with increased respired carbon upwells and mixes with the surface water, resulting in only a moderate change in surface chemistry. Consequently, bottom waters ventilated from the Southern Ocean are less efficient than NADW at storing atmospheric CO2 when the remineralization depth increases.

It is noteworthy that the regions within which changes in the remineralization depth control atmospheric pCO2 are globally distributed. We performed experiments in which we restricted the change in the remineralization depth to individual basins. We found that 38% of the total reduction in atmospheric pCO2 can be attributed to the increase in the remineralization depth. The shift could have been due to changes in circulation and/or changes in biological production and remineralization. If we ignore the possible differences in the circulation and attribute the shift to a deepening of remineralization, perhaps caused by a slowdown of bacterial consumption of sinking organic matter at colder temperatures or increased ballast minerals in a drier climate, the implied change in remineralization depth corresponds to an increase in the e-folding depth from 204 to 313 m (see Supplementary Information).

Palaeoclimate records suggest that respired carbon and nutrients were shifted from intermediate waters towards deep waters in the glacial ocean. The shift could have been due to changes in circulation and/or changes in biological production and remineralization. If we ignore the possible differences in the circulation and attribute the shift to a deepening of remineralization, perhaps caused by a slowdown of bacterial consumption of sinking organic matter at colder temperatures or increased ballast minerals in a drier climate, the implied change in remineralization depth corresponds to an increase in the e-folding depth from 204 to 313 m (see Supplementary Information).
Figure 3 | The DIC response decomposed into soft-tissue, carbonate and gas-exchange pump components. The response of DIC (μmol kg\(^{-1}\)) to an increase in the e-folding depth from 204 to 228 m (the exponent \(b\) changes from 1.0 to 0.9). The total change in DIC (first row) is separated into contributions from the soft-tissue (second row), carbonate (third row) and gas-exchange (fourth row) pumps. a, Zonally averaged fields along the Atlantic Ocean. b, Depth-averaged fields in the global ocean. Warm colours (positive values) represent an increase in DIC (μmol kg\(^{-1}\)). Cold colours (negative values) represent a decrease.
Figure 4 | The alkalinity response decomposed into soft-tissue and carbonate pump components. The response of alkalinity (μeq kg⁻¹) to an increase in the e-folding depth from 204 to 228 m (the exponent b changes from 1.0 to 0.9). The total change in alkalinity (first row) is separated into contributions from the soft-tissue (second row) and carbonate (third row) pumps. Zonally averaged fields along the Atlantic are shown for the nutrient-restoring model and the constant-export model. Warm colours (positive values) represent an increase in alkalinity. Cold colours (negative values) represent a decrease.

Climate-change-induced changes in remineralization depth might also influence the oceanic uptake of anthropogenic carbon in the future. A transient simulation shows that considerable fractions (more than 30%) of the full response occur on timescales of decades, although it takes several thousands of years to reach equilibrium (Supplementary Fig. S3). This indicates that possible changes in remineralization depth could feed back on twenty-first century climate change. How remineralization depth will respond to future climate change is uncertain at present, but our work suggests that the impact on the global carbon cycle could be substantial.

Methods

Global ocean biogeochemistry model. The OCMIP-2 (ocean carbon-cycle model intercomparison project phase 2; ref. 29) ocean biogeochemistry model is coupled with an off-line primitive equation ocean general circulation model. This model uses Newton’s method to solve for steady-state solutions and hence is several orders of magnitude faster than a traditional approach based on time stepping. The time efficiency allows us to perform a systematic analysis of the model for changes in remineralization depth for which the ocean takes several thousand years to reach equilibrium (Supplementary Fig. S3).

In this model, net community production is simulated by restoring model PO₄ towards observations in the top 75 m of the ocean only when model PO₄ exceeds observations. 74% of the net production becomes dissolved organic matter (DOM) whereas the remaining fraction becomes particulate organic matter (POM). DOM is decomposed to inorganic form with a mean lifetime of 1.7 years. The downward flux of POM follows the power law curve as described by ref. 6. Thus the remineralization of POM, \( J(z) \), below the production zone \( (z_c = -75 \text{ m}) \) is represented by

\[
J(z) = \frac{\partial}{\partial z} \left( \frac{z}{z_c} \right)^b F(z_c)
\]

in which \( F(z_c) \) is the amount of POM exported from the production zone. The control value of the exponent \( b = 0.97 \), the value estimated in ref. 8 using the global PO₄, alkalinity and DIC data. The exponent \( b \) can be associated with the e-folding depth, the depth at which the downward flux of POM drops to a factor of 1/e. The corresponding e-folding depths for the exponents \( b = 153 \text{ m} (b = 1.4) \), 162 m \( (b = 1.3) \), 173 m \( (b = 1.2) \), 186 m \( (b = 1.1) \), 204 m \( (b = 1.0) \), 210 m \( (b = 0.97) \), 228 m \( (b = 0.9) \), 262 m \( (b = 0.8) \), 313 m \( (b = 0.7) \), 397 m \( (b = 0.6) \) and 554 m \( (b = 0.5) \).

The CaCO₃ production is proportional to the production of particulate organic carbon with a rain ratio of \( r = 0.08 \). Subsequently CaCO₃ dissolves following an exponential curve with an e-folding length scale of 2,100 m, which is fixed throughout this study. This model assumes constant stoichiometric ratios of C:N:P = 137:16:1. All of these ‘control’ parameter values were determined in ref. 8. There are no sedimentation processes in this model.

Nutrient-restoring model and constant-export model. In the nutrient-restoring model, we follow the nutrient-restoring scheme as implemented in OCMIP-2 (ref. 29) where model-predicted surface nutrients are restored towards observations with a timescale of 30 days only when model PO₄ exceeds the observations. In the constant-export model new production is kept fixed to the value diagnosed from the control run unless the PO₄ concentration in the production zone drops to zero. Where PO₄ is below zero, new production stops.
Preformed and remineralized PO$_4$\textsubscript{3-}. We divide the total PO$_4$ in the ocean into preformed and remineralized components:

$$\text{PO}_4^{3-} = \text{PO}_4^{3-}\text{pre} + \text{PO}_4^{3-}\text{rem}$$

By definition, all PO$_4$ in the surface ocean is preformed and the preformed PO$_4$ in the subsurface ocean is the PO$_4$ that is transported from the surface by ocean circulation. The remineralized PO$_4$ is the PO$_4$ that is added to the subsurface water by remineralization of organic material. Here we compute the subsurface preformed PO$_4$ by integrating the convection over the entire surface area $\Omega$,

$$\text{PO}_4^{3-}\text{pre} = \int_{\Omega} \text{PO}_4(r_i) G(r_i) \, d^2 r_i$$

where PO$_4(r_i)$ is the surface concentration of PO$_4$ and $G(r_i)$ is the volume integrated Green's function denoting the volume of the ocean per unit surface area that is ventilated from the surface grid point, $r_i$ (ref. 18). The remineralized PO$_4$, $\text{PO}_4^{3-}\text{rem}$, is obtained by subtracting the preformed PO$_4$ from the total PO$_4$.

### Carbon pump separation and alkalinity pump separation

We apply the carbon pump separation method\textsuperscript{1} to the DIC response pattern\textsuperscript{4}. Thus the total response of DIC ($\Delta$DIC) becomes

$$\Delta \text{DIC} = \Delta \text{DIC}^{\text{soft}} + \Delta \text{DIC}^{\text{carb}} + \Delta \text{DIC}^{\text{alk}}$$

in which the change in DIC is attributed to the contributions from the soft-tissue pump, the carbonate pump and the gas-exchange pump. The soft-tissue pump component of the DIC response is obtained by multiplying the stoichiometric ratio of C:P by the PO$_4$ response:

$$\Delta \text{DIC}^{\text{soft}} = r_{\text{C:P}} \times \Delta \text{PO}_4$$

(1)

and the carbonate pump component of the DIC response is the nitrate-corrected alkalinity response\textsuperscript{28}:

$$\Delta \text{DIC}^{\text{carb}} = (\text{ALK} + r_{\text{N:PO}4} \times \Delta \text{PO}_4)/2$$

(2)

The gas-exchange component of the DIC response is obtained by subtracting equations (1) and (2) from the total DIC response. Likewise the alkalinity response is separated into the contributions from the soft-tissue pump and the carbonate pump\textsuperscript{2}. Thus the total response of the alkalinity (ALK) becomes

$$\Delta \text{ALK} = \Delta \text{ALK}^{\text{soft}} + \Delta \text{ALK}^{\text{carb}}$$

where the soft-tissue pump component is estimated as

$$\Delta \text{ALK}^{\text{soft}} = -r_{\text{N:PO}4} \times \Delta \text{PO}_4$$

and the remaining $\Delta \text{ALK}$ becomes the carbonate pump component of the alkalinity response.

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Author contributions

E.Y.K. and F.P. initiated the project. E.Y.K. carried out model simulations and analyses with advice from F.P. and J.L.S. E.Y.K. wrote the paper and the Supplementary Information with input from F.P. and J.L.S.

Additional information

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