



## Abstract

The spatial distribution of the air-sea flux of carbon dioxide is a poor indicator of the underlying ocean circulation and of ocean carbon storage. The weak dependence on circulation arises because mixing-driven changes in solubility-driven and biologically-driven air-sea fluxes largely cancel out. This cancellation occurs because mixing driven increases in the poleward residual mean circulation results in more transport of both remineralized nutrients and heat from low to high latitudes. By contrast, increasing vertical mixing decreases the storage associated with both the biological and solubility pumps, as it decreases remineralized carbon storage in the deep ocean and warms the ocean as a whole.

## 1 Introduction

The ocean carbon cycle plays a critical role in setting today's climate by determining the partitioning of carbon between the atmosphere and ocean. Two natural ways of characterizing this cycle are the meridional transport and the vertical profile of carbon within the ocean. The north-south gradient of the meridional transport of carbon is primarily balanced by air-sea fluxes of carbon dioxide, leading to the obvious question of whether measuring such fluxes (e.g., Takahashi et al., 2009) would provide important constraints on ocean circulation or carbon storage. Using a suite of ocean circulation models with different parameterizations of interior mixing, we show that the air-sea CO<sub>2</sub> flux is insensitive to the circulation, in contrast to the storage of carbon by the ocean.

We analyze this contrasting behavior in terms of the oceanic solubility and biological pumps of carbon. Cold high latitude surface waters hold more dissolved inorganic carbon (DIC) than warm, low-latitude waters. The solubility pump is the result of processes whereby these cold waters sink, injecting CO<sub>2</sub> into the deep ocean (Volk and Hoffert, 1985). The biological pump is the surface to deep gradient in DIC associated

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with the production of planktonic organic matter at the ocean surface and the transport and subsequent decomposition of biogenic material throughout the water column.

The extent to which changes in ocean ventilation will impact the natural carbon pumps and air-sea CO<sub>2</sub> exchange is a topic of current research and debate. In an intriguing new study Le Quere et al. (2007) suggest that changes in Southern Ocean ventilation associated with increased Westerlies will result in a loss of biologically sequestered CO<sub>2</sub> from the deep ocean via increased Southern Ocean upwelling and a general weakening of the natural biological pump in a future warmer climate. Changes in the rate of convection and deep water formation in the Southern Ocean have been shown to strongly impact the oceanic biological pump and atmospheric pCO<sub>2</sub> (e.g. Marinov et al., 2008a, b) and are likely a main driver for glacial-interglacial changes in atmospheric pCO<sub>2</sub> (e.g., Sarmiento and Toggweiler, 1984; Anderson et al., 2007; Toggweiler et al., 2006).

In this paper, we use a suite of models in which vertical and lateral mixing are varied, producing a range of ocean circulation. We show that while these changes produce compensating responses in the air-sea fluxes associated with the solubility and biological pumps, they do not produce a compensation in the inventories of carbon associated with these pumps. As a result, air-sea CO<sub>2</sub> flux is decoupled both from ocean circulation and from the storage of carbon within the ocean.

## 2 Methods

We use the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM) version 3, previously used for a number of biogeochemical dynamics studies (Gnanadesikan et al., 2002, 2004; Marinov et al., 2006, 2008a, b). The biogeochemistry component follows OCMIP-2 specifications, with new biological production calculated by restoring our limiting nutrient PO<sub>4</sub> to observations PO<sub>4,obs</sub> at each time step  $t$ , whenever PO<sub>4</sub> > PO<sub>4,obs</sub>:

$$J_{\text{prod}}(x, y, z, t) = (\text{PO}_4(x, y, z, t) - \text{PO}_{4,\text{obs}}(x, y, z, t)) / \tau \text{ for } z < 75 \text{ m} \quad (1)$$

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where  $\tau$  is the biological time scale,  $x, y, z$  are model longitude, latitude and depth. Globally integrated  $\text{PO}_4$  is set constant in all experiments.

Four different versions of MOM3 are analyzed here.

- Our “Control” stands for low  $A_i$  and low  $K_v$  model. Diapycnal mixing  $K_v$  varies hyperbolically from  $0.15 \text{ cm}^2/\text{s}$  at the surface to  $1.3 \text{ cm}^2/\text{s}$  at 5000 m, and isopycnal mixing  $A_i$  is set to  $1000 \text{ m}^2/\text{s}$  everywhere in the ocean.
- In the high  $K_v$  model,  $K_v$  varies hyperbolically from  $0.6 \text{ cm}^2/\text{s}$  at the surface to  $1.3 \text{ cm}^2/\text{s}$  at 5000 m everywhere in the ocean, while  $A_i$  is as in the Control model.
- In the high  $A_i$  model  $K_v$  is as in the Control model, but  $A_i$  is set to  $2000 \text{ m}^2/\text{s}$ .
- In the high  $A_i$ -high  $K_v$  model,  $A_i$  is  $2000 \text{ m}^2/\text{s}$  and  $K_v$  is as in the high  $K_v$  model.

Each of these models is run in three configurations that allow us to isolate the impacts of the biological and solubility effects on the carbon pump. In the ABIOTIC run biology is turned off (Murnane et al., 1999), such that the resulting  $\text{CO}_2$  distribution and air-sea  $\text{CO}_2$  fluxes depend only on temperature and salinity. In the BIOTIC run we turn the solubility pump off by setting the ocean temperature and salinity constant everywhere at the surface in the calculation of the air-sea  $\text{CO}_2$  exchange calculation. Finally, in the FULL configuration we turn on biology such that the resulting  $\text{CO}_2$  distributions and fluxes depend on both biological and solubility effects.

### 3 Results and discussion

Subgridscale mixing and resulting changes in ocean circulation have a major impact on the natural carbon pumps and on the air-sea  $\text{CO}_2$  fluxes. Figure 1 shows results from our 4 models with different mixings. Each of these models was run in the ABIOTIC, BIOTIC and FULL configurations. The surface to deep gradients in DIC in each of these

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configurations are representative for the the strength of the solubility pump (Fig. 1a), the biological pump (Fig. 1d) and the full carbon pump (Fig. 1g) at equilibrium. Figure 1g supports out contention that the separation of total DIC and total carbon pumps into components works well, at least in the global average. Figure 1 also shows the abiotic, biological and total air-sea CO<sub>2</sub> fluxes (middle column) and transports (right hand column) corresponding to the pumps in the left hand column.

Two important observations are:

- I. an increase in  $K_V$  decreases the ocean carbon storage associated with both biological and solubility carbon pumps.
- II. an increase in  $K_V$  increases the magnitude of both the biological and the abiotic air-sea fluxes. The two effects cancel each other, such that the total air-sea CO<sub>2</sub> flux changes little.

The impact of changing diapycnal and isopycnal mixing on oceanic circulation and oceanic heat transport is summarized in Fig. 2 (see also Gnanadesikan (1999, 2003) and Marinov et al., 2008a). Increasing vertical mixing  $K_V$  affects oceanic circulation in 2 major ways:

- A. Increasing the relative contribution of the Southern Ocean waters to the deep ocean via increased Southern Ocean convection and AABW formation (blue arrow in Fig. 2) and the upwelling of high-nutrient deep waters in the tropics ( $T_U$  in Fig. 2) which is balanced by increased downward diffusion of heat.
- B. changing the residual mean circulation ( $T_S$  in Fig. 2). In detail, increasing  $K_V$  increases the conversion from dense to light water in the tropics and the thermocline depth. This in turn increases the southward eddy transport  $T_{\text{eddy}}$  and the Northern Hemisphere overturning  $T_n$  and thus increases poleward water and heat transport in both hemispheres.

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We will show below that mechanism A explains most of the observed carbon pump differences among models (Observation I above), while mechanism B dominates the air-sea flux differences among models (Observation II above).

5 Everywhere in the ocean our biologically limiting nutrient is the sum between a preformed and a remineralized component. Preformed nutrients are the nutrients that sink into the ocean interior without having been utilized in photosynthetic production at the ocean surface; they signal an inefficient biological pump (Ito and Follows, 2005; Marinov et al., 2006). Remineralized nutrients are those respired throughout the water column; these are stoichiometrically associated with remineralized carbon. The deep  
10 ocean is ventilated by the Southern Ocean and the NADW. The Southern Ocean has relatively higher preformed nutrients (i.e., a more inefficient biological pump) and lower SST.

Increasing  $K_V$  makes the deep ocean look more like the Southern Ocean, increasing preformed nutrients and decreasing temperature (Fig. 3a) along the AABW pathway and in the deep. The consequences are a “leakier” or less efficient biological carbon pump in the high  $K_V$  model compared to the control LL model (see also Marinov et al., 2008a) and an increase in the solubility driven (abiotic) DIC in the AABW and below 3000 m (Fig. 3c). For a constant total phosphate, an increase in preformed nutrient will correspond to a decrease in remineralized nutrient along the AABW pathway  
20 and in the deep ocean (Fig. 3d). That these changes in remineralized nutrient are reflected in biological carbon is shown in Fig. 3b (compare Figs. 3b and 3d) as well as in Fig. 1d, where the intermodel differences in the biological pump as reflected by the DIC gradient between the surface and deep ocean are shown to be driven by changes in remineralized nutrients.

25 The increase in vertical mixing  $K_V$  in low latitudes also results in deeper penetration of surface heat, increasing temperatures above 2500 m depth. This warming translates into less net storage of abiotic DIC in the deep ocean (Fig. 3a, c), and thus a less efficient solubility pump in the high  $K_V$  compared to the control model (Fig. 1a).

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Because of its lower total remineralized inventory and overall warmer temperatures, the high  $K_v$  model stores less carbon than the control model in the ocean. Overall, increased Southern Ocean transformation and low latitude mixing (Mechanism A) result in decreases in the efficiency of both biological and solubility pumps.

But what are the implications for air-sea fluxes and the meridional transport of carbon in the ocean? And how are air-sea  $\text{CO}_2$  fluxes connected to the carbon pumps? The net air-sea  $\text{CO}_2$  exchange reflects a competition between biological and solubility effects, with the biological air-sea  $\text{CO}_2$  flux opposing the solubility air-sea  $\text{CO}_2$  flux throughout most of the ocean. The abiotic effect wins over the biological effect such that on average high latitudes are a sink for atmospheric  $\text{CO}_2$  (Fig. 1b, e, h). Interestingly, this is very different from the total carbon pump (reflected by the DIC surface-to-deep gradients) in the ocean, which is dominated by the biological rather than the solubility pump (Fig. 1a, d, g).

The abiotic air-sea  $\text{CO}_2$  flux follows to first order the inverse of the heat flux (Keeling et al., 1993). Upwelling brings cold water to the surface at the Equator where it is heated up; the decrease in solubility results in loss of carbon to the atmosphere. As this water moves at the surface to higher latitudes in both hemispheres (e.g., through currents such as Gulf Stream, Kuroshio or through eddy diffusive transport) it gradually loses heat and gains  $\text{CO}_2$ ; low latitude outgassing is offset by  $\text{CO}_2$  ingassing in high latitudes.

The biological  $\text{CO}_2$  flux reflects the preformed nutrient distribution at the ocean surface. In high latitude regions such as the Southern Ocean, remineralized nutrients and carbon are upwelled to the surface. Since biological uptake is inefficient here relative to the upward supply, preformed nutrients and  $p\text{CO}_2$  at the ocean surface build up, resulting in  $\text{CO}_2$  outgassing to the atmosphere. Technically, the rate of biological outgassing to the atmosphere is controlled by the efficiency of conversion of deep remineralized nutrients to surface preformed nutrients; more surface preformed nutrients signal more inefficient biology and more biological outgassing.

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At steady state, large biological uptake in the subtropics is required to counterbalance high latitude biological degassing. Biological uptake is generally efficient in the subtropics; this translates into near-zero surface preformed nutrients and net uptake of carbon from the atmosphere. High export of organic matter at low latitudes is reflected by high remineralized nutrients and carbon right below the surface (with highest concentrations above 1000 m depth). From a nutrient budgeting perspective, we see a net conversion of preformed surface nutrients to remineralized sub-surface nutrients, and this conversion controls the rate of biological carbon ingassing into the tropical ocean.

In order to understand the impact of mixing on the air-sea carbon exchange, it is easiest to think of the abiotic and biological air-sea carbon fluxes as the horizontal gradients of the abiotic and biological transports of carbon, respectively (Fig. 1c, f). The depth integrated transport of abiotic carbon is a consequence of the net poleward heat transport in each hemisphere, while the biological meridional transport of carbon is given to a first order by the poleward transport of remineralized nutrients, adjusted by the appropriate stoichiometric constant.

Residual mean circulation is a sum between northwards Ekman advection and counteracting southward eddy transport. Changes in the residual mean circulation due to increased diapycnal mixing create a pattern in remineralized nutrients analogous to the temperature pattern (Fig. 3a, b), with increased values in sub-surface low latitudes and decreased Southern Ocean and deep ocean values. Increased diapycnal mixing increases the poleward residual mean circulation  $T_{res}$ , transporting poleward both higher remineralized nutrients (and thus biological DIC) and warmer tropical waters (and thus more heat). The resulting increase in poleward transport of biological DIC compensates almost exactly the increase in equatorward solubility driven (abiotic) DIC, such that total carbon transport and resulting air-sea  $\text{CO}_2$  fluxes do not change much with mixing (Fig. 1h, i).

Because biological and abiotic  $\text{CO}_2$  fluxes (and the corresponding transports) work in opposition throughout most of the ocean, surface  $\text{CO}_2$  levels are set by a “compensation” mechanism between biological and solubility effects. We have seen that while

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both of these fluxes vary strongly with mixing, the total air-sea CO<sub>2</sub> flux and carbon transport vary much less. This can be explained by a scaling of the “compensation” mechanism with mixing, i.e., the change in the biological flux cancels to a first order the corresponding change in the abiotic flux.

Vertical diffusion increases the poleward heat transport by 0.8 PW in the Southern Hemisphere and 0.2 PW in the Northern Hemisphere (Gnanadesikan et al. 2002). Lateral diffusion has a smaller impact, increasing poleward heat transport in the Southern Hemisphere by 0.4 PW and reducing it in the Northern Hemisphere. Interestingly, these responses are twice as strong in the Southern compared to the Northern Hemisphere. Similarly, the sensitivity of both abiotic and biological fluxes to changes in subgridscale mixing or wind intensity is highest in the Southern Ocean. This is consistent with the results of Sarmiento, et al. (1998) who find large sensitivity in the Southern Ocean biological and abiotic fluxes to changes in stratification following climate change. Watson and Orr (2003) analyze a set of models with identical (OCMIP) biogeochemistry but different model physics and different boundary conditions, and find largest disagreement between biological, solubility and anthropogenic CO<sub>2</sub> air-sea fluxes in the Southern Ocean.

Finally, we note that isopycnal mixing  $A_i$  has a much smaller effect on carbon pumps and air-sea fluxes than diapycnal mixing  $K_v$ . An increase in  $A_i$  decreases on average the large scale overturning circulation, decreases equatorial upwelling and inhibits the large oceanic cooling associated with the Western Boundary Current. Therefore, changes in the abiotic air-sea CO<sub>2</sub> flux due to increasing  $A_i$  can locally oppose changes in fluxes due to increasing  $K_v$ .

## 4 Conclusions

We have shown that increasing mixing decreases both the solubility and the biological carbon pumps and carbon storage in the ocean but has little impact on the total air-sea CO<sub>2</sub> flux. The difference arises because different parts of the ocean circulation

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are responsible for setting the magnitude of the pumps and fluxes. AABW formation plays a critical role in setting the inventory of carbon, and because it involves relatively small changes in temperatures does not imply a compensation between the biological and solubility pumps. By contrast, air-sea CO<sub>2</sub> fluxes depend on the the transport of warm waters with higher remineralized nutrients from low to high latitudes. Changes in the solubility CO<sub>2</sub> flux with vertical mixing are explained by underlying changes in temperature and meridional heat transport, while changes in the biological CO<sub>2</sub> flux with vertical mixing mirror changes in remineralized nutrients and their meridional transport. As a result, we observe a strong compensation between changes in the biological and solubility CO<sub>2</sub> fluxes and transports such that the total flux and the corresponding total meridional transport remain largely unchanged (Fig. 1).

If the compensation mechanism were to hold true in the real ocean, potential changes in circulation due to to climate change, while strongly affecting the individual pump strength and the natural carbon sequestration in the deep ocean, would not have an observable impact on the total CO<sub>2</sub> fluxes and the meridional transport of natural carbon.

Furthermore, research has suggested that SST differences between GCMs forced with different winds are amplified in coupled simulations compared to ocean only models such as ours where atmospheric temperature is held fixed (Gnanadesikan and Anderson, 2009). We might therefore expect an even larger response of abiotic and biological CO<sub>2</sub> fluxes and carbon pumps to winds or changes in mixing in coupled models. The existence of this compensation mechanism might be even more important and needs to be further tested in such coupled setups.

Our analysis highlights the critical importance of boundary conditions and the parameterization of heat fluxes and biological production at the ocean surface in determining air-sea CO<sub>2</sub> fluxes and oceanic carbon pump strengths. Important caveats in our model are (a) surface restoring of temperature and salinity, which constraint strongly the solubility air-sea carbon flux and carbon pump, and (b) surface restoring of nutrients and a simplistic parametrization of biological production, which constraint the

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biological air-sea flux and biological carbon pump. Would the compensation mechanism presented here still hold up in the context of a coupled model which allows boundary conditions to change freely? A model with a more complex representation of production and heat fluxes might allow the surface temperature and nutrients to vary more significantly in response to changes in vertical mixing, potentially affecting the air-sea heat and carbon fluxes. Such feedbacks between changes in mixing and biogeochemistry are constrained in our model and need to be further explored.

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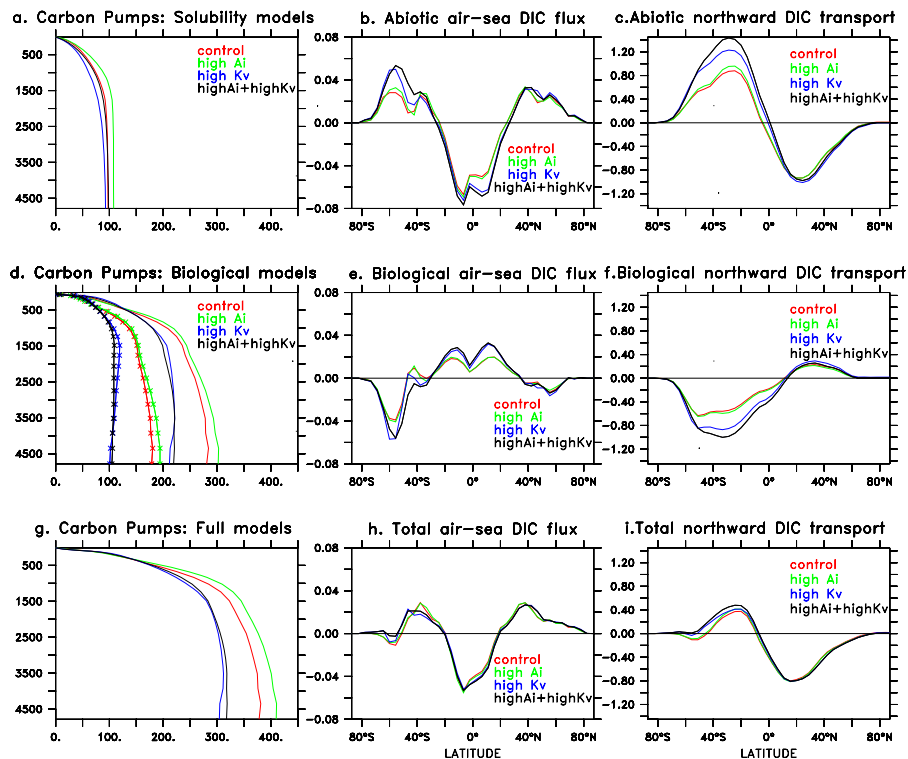
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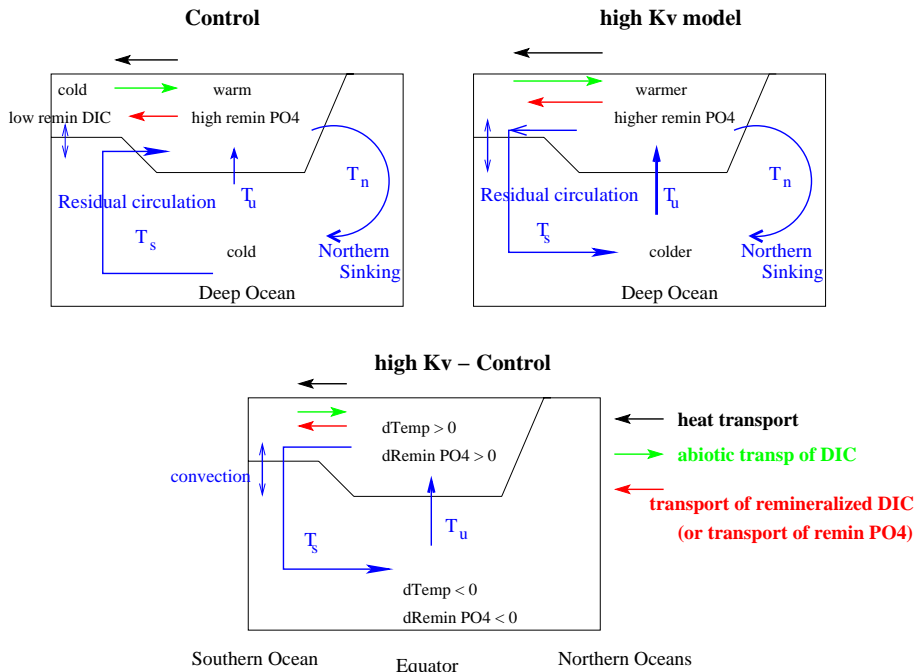
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**Fig. 1.** Pumps, carbon fluxes and transports in 4 models with different mixing. Abiotic carbon pump, air-sea DIC flux in Pg C/yr/degree and northward DIC transport integrated over the entire watercolumn in Pg C/yr (top panels, **a–c**). Biology only carbon pump, air-sea DIC flux and transport (middle panels, **d–f**). Total carbon pump, air-sea DIC flux and transports from a model that includes both biotic and solubility pumps (bottom three panels, **g–i**). Fluxes and transports are zonal and annual mean averages. Pumps represent the zonally and meridionally integrated DIC at each depth minus average surface DIC in  $\mu\text{mol/kg}$ . For each model, the FULL carbon pump (panel g) is the sum of the ABIOTIC (panel a) and BIOLOGICAL (panel d) carbon pumps. Intermodel differences in biological carbon pumps (unstarred lines in panel d) can be explained by intermodel differences in remineralized nutrients (starred lines in panel d show remineralized nutrients times a C:P ratio of 117).

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**Fig. 2.** Schematic showing the impact of increased diapycnal mixing on circulation and the transport of abiotic DIC and remineralized PO<sub>4</sub> in our GCM. Changes in the abiotic DIC transport and biological DIC transport cancel each other out. The total (abiotic+biological) DIC transport is similar in the control and high  $K_v$  models and is northward.

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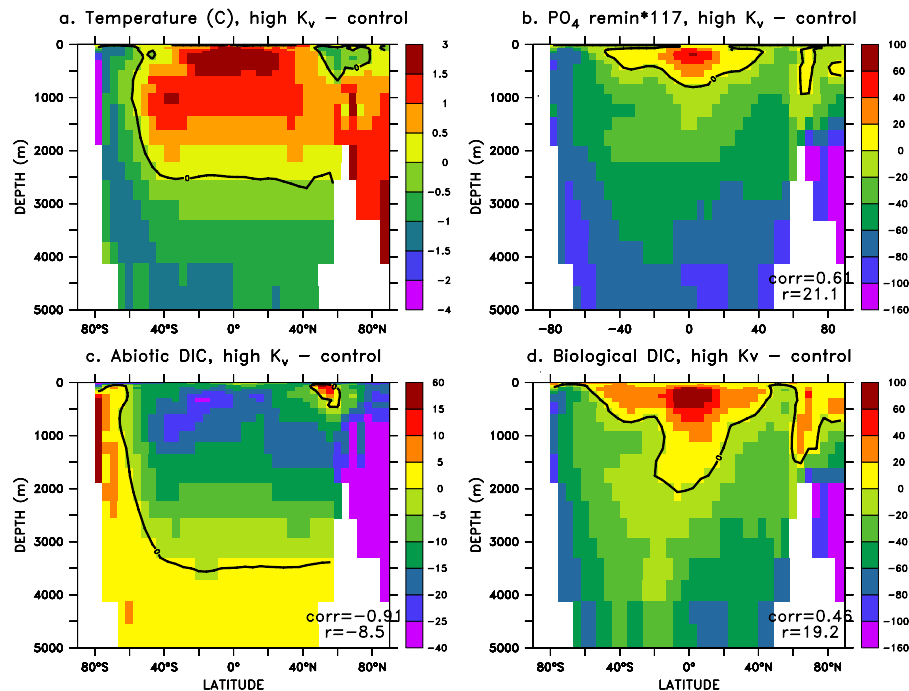
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**Fig. 3.** Zonally averaged annual mean differences between the high  $K_v$  and control models: (a) Temperature ( $^{\circ}\text{C}$ ) (b) Remineralized DIC ( $\mu\text{mol}/\text{kg}$ ) calculated as remineralized  $\text{PO}_4$  (model tracer) times 117. (c) salinity normalized DIC ( $\mu\text{mol}/\text{kg}$ ) from the ABIOTIC run, (d) biological DIC calculated from BIOTIC runs ( $\mu\text{mol}/\text{kg}$ ). Small differences between b and d are due to intermodel differences in air-sea surface  $\text{CO}_2$  disequilibrium. Correlations (corr) and regression coefficients ( $r$ ) with the temperature difference in panel a are shown in each case. Globally, remineralized DIC is well correlated to temperature (corr=0.61). Abiotic DIC differences are due primarily to intermodel differences in temperature (corr=-0.91), and to a lesser extent to intermodel salinity differences.

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