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Mechanisms governing interannual variability in upper-ocean inorganic carbon system and air–sea CO_2 fluxes: Physical climate and atmospheric dust

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ABSTRACT

We quantify the mechanisms governing interannual variability in the global, upper-ocean inorganic carbon system using a hindcast simulation (1979–2004) of an ecosystem-biogeochemistry model forced with time-evolving atmospheric physics and dust deposition. We analyze the variability of three key, interrelated metrics—air–sea CO_2 flux, surface-water carbon dioxide partial pressure pCO_2, and upper-ocean dissolved inorganic carbon (DIC) inventory—presenting for each metric global spatial maps of the root mean square (rms) of anomalies from a model monthly climatology. The contribution of specific driving factors is diagnosed using Taylor expansions and linear regression analysis. The major regions of variability occur in the Southern Ocean, tropical Indo-Pacific, and Northern Hemisphere temperate and subpolar latitudes. Ocean circulation is the dominant factor driving variability over most of the ocean, modulating surface dissolved inorganic carbon that in turn alters surface-water pCO_2 and air–sea CO_2 flux variability (global integrated anomaly rms of 0.34 Pg C yr^{-1}). Biological export and thermal solubility effects partially damp circulation-driven pCO_2 variability in the tropics, while in the subtropics, thermal solubility contributes positively to surface-water pCO_2 and air–sea CO_2 flux variability. Gas transfer and net freshwater inputs induce variability in the air–sea CO_2 flux in some specific regions. A component of air–sea CO_2 flux variability (global integrated anomaly rms of 0.14 Pg C yr^{-1}) arises from variations in biological export production induced by variations in atmospheric iron deposition downwind of dust source regions. Beginning in the mid-1990s, reduced global dust deposition generates increased air–sea CO_2 outgassing in the Southern Ocean, consistent with trends derived from atmospheric CO_2 inversions.

1. Introduction

The ocean exhibits variability in physical circulation on sub-annual to decadal and longer time-scales that in turn drives substantial changes in regional to basin-scale biogeochemistry and air–sea CO_2 fluxes (Chavez et al., 1999; Le Quéré et al., 2000, 2003; Gruber et al., 2003; Dore et al., 2003; Bates et al., 2003; Corbiere et al. 2007). Variations in aeolian dust deposition may generate additional biogeochemical variability in iron-limited ocean regions (Jickells et al., 2005; Patra et al., 2007; Cassar et al., 2007; Tagliabue et al., 2008; Aumont et al., 2008). Such variability offers unique, natural perturbation "experiments" for probing underlying mechanisms and characterizing potential responses to and feedbacks on natural and anthropogenic climate change (Boyd and Doney, 2002). Variability in air–sea CO_2 fluxes also imprints on the overlying atmosphere (Nevison et al., 2008), and thus directly impacts on so-called "top-down" efforts to reconstruct oceanic and terrestrial carbon sources and sinks by inverting the time/space evolution of atmospheric CO_2 (e.g., Bousquet et al., 2000; Gurney et al., 2002; Peylin et al., 2005).

The amplitude, patterns and phasing of ocean interannual variability are modulated primarily by the major atmosphere and atmosphere–ocean climate modes (Wang and Schimel, 2003). Globally, the largest marine physical and biogeochemical signals arise in the tropical Pacific, modulated by the El Nino-Southern Oscillation (ENSO). The mechanisms of the ENSO biogeochemical response are reasonably well understood (Chavez et al., 1999; Behrenfeld et al., 2006). Regional changes in atmospheric

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convection and trade winds affect upwelling of subsurface water containing high dissolved inorganic carbon (DIC) and carbon dioxide partial pressure (pCO_2) while remotely forced Kelvin waves influence the depth of the thermocline and thus the biogeochemical concentrations of the source waters (Feely et al., 1999). ENSO-related variability also extends over much of the globe because of ocean wave propagation from the tropical Pacific and atmospheric teleconnections (Wang and Schimel, 2003).

Repeat spatial surveys of the air–sea CO_2 partial pressure difference, ΔpCO_2, document a strong correlation between depressed CO_2 efflux to the atmosphere and the onset of El Nino events; observations of La Nina events show a corresponding enhancement of CO_2 efflux (Feely et al., 1999, 2002). Field-based estimates of air–sea CO_2 flux interannual variability are about $\pm 0.2\,Pg\,C\,yr^{-1}$ (Feely et al., 2002). Ocean models suggest comparable, or somewhat lower, estimates of the interannual variability of ± 0.13 to $\pm 0.3\,Pg\,C\,yr^{-1}$, depending in part on the spatial and temporal analysis domains (Le Quéré et al., 2000; Obata and Kitamura, 2003; McKinley et al., 2004; Wetzel et al., 2005).

In the extratropics three major climate modes cause interannual variability, the North Atlantic Oscillation (NAO) (Visbeck et al., 2001; Gruber et al., 2003; Bates et al., 2003), Pacific Decadal Oscillation (PDO) (Takahashi et al., 2003; Feely et al., 2006), and the Southern Annular Mode (SAM) (Lenton and Matear, 2007; Lovenduski et al., 2007; Le Quéré et al., 2007; Verdy et al., 2007). All three modes involve atmospheric pressure oscillations that drive substantial changes in the strength and location of the surface winds, ocean upwelling, ocean convection patterns, sea-surface temperature (SST), and air–sea heat and freshwater fluxes on regional scales. The impact on air–sea CO_2 fluxes depends upon the interaction of several, often competing, climatic factors such as thermal solubility (SSTs), biological drawdown of DIC, upwelling/mixing of nutrient- and DIC-rich waters, net surface freshwater fluxes (through dilution of *DIC* and alkalinity, *Alk*), and wind speed.

Using ocean time-series data from Bermuda, Gruber et al. (2003) suggest a correlation of negative NAO index years with deeper mixed layers, lower SSTs, increased entrainment and biological production, and enhanced CO_2 uptake. Extrapolating from Bermuda to the entire subtropical gyre leads to an interannual air–sea CO_2 flux variability of $\pm 0.2\,Pg\,C\,yr^{-1}$; assuming that the subtropic and subpolar air–sea CO_2 flux variability is in phase, as done by Gruber et al. (2003), increases the variability of the whole North Atlantic to $\pm 0.3\,Pg\,C\,yr^{-1}$. From an extended Bermuda time-series, Bates (2007) estimates subtropical interannual air–sea CO_2 flux variability of $0.2–0.3\,Pg\,C\,yr^{-1}$. However, as discussed below there is reason to question the assumptions that the North Atlantic basin as a whole or even just the subtropics vary coherently, suggesting that these extrapolations from the Bermuda record may overestimate basin-integrated interannual variability.

Model simulations indicate considerably weaker Northern Hemisphere extratropical variability (Le Quéré et al., 2000, 2003). Obata and Kitamura (2003), for example, predict that the interannual variability in North Atlantic air–sea CO_2 flux is only $\pm 0.04\,Pg\,C\,yr^{-1}$ in the subtropics and $\pm 0.03\,Pg\,C\,yr^{-1}$ in the subpolar gyre. McKinley et al. (2006) analyze tropical and North Pacific CO_2 variability from seven biogeochemical models. They report relatively weak interannual variability in North Pacific air–sea CO_2 flux of ± 0.03 to $\pm 0.11\,Pg\,C\,yr^{-1}$ for the basin, reflecting in part cancellation of out-of-phase regional air–sea flux anomalies. Although the model air–sea CO_2 fluxes are correlated with the Pacific Decadal Oscillation, the total projection on the PDO is small because PDO-generated surface temperature, DIC and alkalinity anomalies have opposing effects on surface pCO_2.

The Southern Annular Mode is the dominant climate mode over the Southern Ocean and is approximately zonally symmetric and circumpolar about the Antarctic continent. A positive SAM phase is associated with increased surface wind stress, enhanced upwelling of carbon rich, subsurface circumpolar deep water, and efflux of ocean CO_2 to the atmosphere (Lenton and Matear, 2007; Lovenduski et al., 2007; Le Quéré et al., 2007). Simulated air–sea CO_2 flux variability in the Southern Ocean ranges from ± 0.07 (Obata and Kitamura, 2003) to $\pm 0.19\,Pg\,C\,yr^{-1}$ (Lovenduski et al., 2007).

Historical ocean carbon data are too sparse, except for a few regions, to fully resolve upper-ocean carbon system and air–sea CO_2 flux variability on the required regional and monthly scales (Bender et al., 2002). This will likely remain true in the near-term on a global-scale, even with the recent growth in instrumented biogeochemical moorings and volunteer observing ship (VOS) pCO_2 transects (e.g., Metzl et al., 2007; Doney et al., 2009b). Further, while critical, the data from observational networks often cannot differentiate clearly among different potential factors driving the observed variability.

Numerical simulations provide important tools to aid the interpretation of observational data, extrapolate/interpolate air–sea CO_2 fluxes to the required regional and monthly scales in a dynamically consistent manner, and characterize the underlying physical–biogeochemical processes (e.g., Le Quéré et al., 2000; Obata and Kitamura, 2003; McKinley et al., 2004, 2006; Wetzel et al., 2005). Here we present a globally consistent analysis of upper-ocean biogeochemical interannual variability from a numerical hindcast (1979–2004) that exhibits good skill relative to observations. The analysis includes both air–sea CO_2 fluxes and surface-water chemistry. We also quantify and partition the underlying forcing factors including atmospheric physical forcing, dust deposition, and ocean circulation and biology.

2. Model formulation

The Community Climate System Model (CCSM-3) ocean Biogeochemical Elemental Cycle (BEC) model consists of an upper-ocean ecological module (Moore et al., 2004) and a full-depth ocean biogeochemistry module (Doney et al., 2006) both embedded in a three-dimensional (3-D) global physical ocean general circulation model (Smith and Gent, 2004; Collins et al., 2006). The model is forced with physical climate forcing from atmospheric reanalysis and satellite data products (Doney et al., 1998, 2007) and time-varying dust deposition (Mahowald et al., 2003). The CCSM BEC model is cast as a set of 3-D, time-varying advection diffusion equations for a suite of tracers $\varkappa$:

$$\frac{\partial \chi}{\partial t} + \nabla \cdot (\vec{u}\chi) - \nabla \cdot (K \nabla \chi) = RHS_{bio} \tag{1}$$

where the second and third terms on the left-hand side are the advective and diffusive divergence terms, respectively, and ∇ is the 3-D del operator. The physical transport is partitioned into resolved advection and parameterized eddy mixing terms; all of the ecological–biogeochemical source/sink terms and surface and sediment fluxes are grouped into the right-hand-side term RHS_{bio}.

2.1. Ecosystem-biogeochemistry modules

The ecosystem module builds on traditional phytoplankton–zooplankton–detritus–nutrient food web models and incorporates multi-nutrient limitation (N, P, Si, Fe) on phytoplankton growth and specific phytoplankton functional groups (Moore et al., 2004). The ecosystem module is coupled with an ocean biogeochemistry module with full carbonate system thermodynamics and air–sea