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Modeling Carbon-Climate Interactions

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How fast greenhouse warming will proceed depends on a large part on how fast carbon dioxide is increasing in the atmosphere. The abundance and growth rate of atmospheric CO_2 is determined not only by the rate of anthropogenic emissions but also by the rates of biogeochemical processes that sequester carbon in the land and ocean. The biogeochemical processes in turn responds to and alters climate and circulation changes. A goal of carbon cycle modeling is to understand the interactions among the physical and biogeochemical processes that exchange carbon among the mobile reservoirs on seasonal to millennial time scales, and to project the future co-evolution of CO_2 and climate. The task requires the synthesis of atmospheric, oceanic and terrestrial observations of carbon inventories and carbon fluxes, and the coupling of the physical and biogeochemical systems.

CO_2 is radiatively active but chemically inert in the atmosphere. The geographic and intraseasonal variations are observed to be small (<10%) and are radiatively insignificant compared to the global annual mean concentrations (~350 ppmv). In this way CO_2 departures from the global mean (referred to as "tracer CO_2") are unimportant for the radiative and biogeochemical processes, and can be treated as an inert tracer in the atmosphere. This has permitted the traditional separation of climate models and carbon cycle models. In climate models, the global annual mean CO_2 concentration is specified. In carbon cycle models, the only crucial atmospheric and oceanic processes to model are those related to large-scale advection and turbulent mixing; the challenge lies in the prescription of source/sink functions for the tracer.

In this paper, we review the status of carbon cycle modeling, and discuss the outlooks for global environment modeling, both in the future and of the future.

MODELING THE CARBON CYCLE

The most simplistic model of the carbon cycle consists of three major carbon reservoirs: the atmosphere (a), terrestrial biosphere (b) and the oceans (o). We shall use the terminology F_{ij} to denote fluxes (Pg C/y) from reservoir i to reservoir j, C_i to denote the carbon inventory (PgC) in reservoir I, and τ_i to denote the turnover times of carbon in the reservoirs. The conservation equations for each reservoir are expressed as:

$$\frac{\partial Ca}{\partial t} + \mathcal{T}_a(Ca) = \text{Anthropogenic sources} - Fab + Fba - Fao + Foa \quad (1)$$

$$\frac{\partial Cb}{\partial t} = Fab - Fba = Fab - \frac{Cb}{\tau b} \quad (2)$$

$$\frac{\partial Co}{\partial t} + \mathcal{T}_o(Co) = Fao - Foa = Fao - \frac{Co}{\tau o} \quad (3)$$

where $\mathcal{T}_a$ and $\mathcal{T}_o$ are the transport operators for the atmosphere and ocean, respectively.

Fab: The terrestrial biosphere absorbs CO_2 through stomates on leaves via photosynthesis. The rate of this absorption is tightly coupled to the energy and hydrologic cycle (see Dickinson, this conference) as well as to the nutrient and biogeochemical status of the soils. The validation of the modeling of *Fab* can be done via comparison to measurements of net primary productivity (NPP) at ecological measurement sites, as well as via comparison to spatial and temporal patterns of the satellite-derived Normalized Difference Vegetation Index (NDVI) globally (e.g. Field *et al*., 1995).

Fba: The return of CO_2 to the atmosphere is more complicated and is tied, most importantly, to the decomposition of plant detritus. The allocation of the photosynthate to different plant compartments (leaves, roots, wood) with different lifetimes implies different composition and turnover times of the dead carbon pools. The decomposition rates are sensitive to temperature, moisture and biogeochemical status of the soils. The turnover time of carbon in (live) vegetation ranges from 1 year for leaves and roots to >30 years for trees; while that in the soils are longer—from ~1 year for herbaceous detritus in the tropics to >200 years for the tundra. Disturbances such as fires and/or deforestation/reforestation further alter the rates of carbon uptake and release. Validation of the respiratory fluxes of CO_2 is difficult. An equilibrium constraint for the pre-industrial demands that the global integral of the respiratory flux to equal the global NPP. The ability of the modeled *Fab* and *Fba* to explain the seasonal cycle of atmospheric CO_2 in the northern hemisphere provides some confidence on the regional aspects of the modeled fluxes.

Fao and *Foa*: Atmosphere-ocean exchange of CO_2 is driven by the difference in CO_2 partial pressures across the air-sea interface. This exchange differs in significant ways from atmosphere-biosphere exchange of CO_2. Firstly, because of the ocean circulation, equilibrium in air-sea exchange is obtained only globally: CO_2 is outgassed from the equatorial oceans and absorbed by the middle-to-high latitude oceans. Secondly, CO_2 comprises ~1% of the dissolved inorganic carbon (DIC) in the surface ocean, with the exact fraction determined by dissociation constants which vary with temperature, salinity and pH. This is often referred to as the “solubility pump” and is dependent on the hydrography of the surface ocean. DIC, in turn, is closely tied to the ocean circulation and the

circulation-driven biological cycle (the biological pump). Photosynthesis takes place where upwelling supplies nutrients to the surface where there is sunlight. The biological processes transfer DIC into particulate organic and inorganic carbon (POC and PIC), which sink and remineralize at depth. Subsequent upwelling returns the carbon to the surface as DIC. Thus, unlike the terrestrial biosphere whose functioning is tied to the near-surface atmospheric state, modeling air-sea CO_2 exchange requires information about the horizontal and vertical exchanges of water masses in the global ocean. The residence time of carbon in the ocean is long ~10^2 to 10^3 years. The biological pump is complex; the POC and PIC reservoirs are small (~1–2 PgC) and highly variable temporally and spatially compared to DIC (~700 PgC); and the residence time of POC and PIC in the surface ocean is short (~days to weeks). Thus its modeling and validation is difficult. Satellite observations (CZCS and SeaWiFS) provide partial validation of the spatial patterns of the modeled surface NPP fields. Validation of the sub-surface parts of the biological pump is difficult.

Models of the dynamics of the terrestrial and oceanic reservoirs are many and of varying complexity (e.g. Sellers *et al*., 1996, 1997; Randerson *et al*., 1997; Doney *et al*., 1999). Forward modeling of atmospheric CO_2 would use fluxes calculated by these biogeochemical models as forcing functions in a global three-dimensional (3D) atmospheric circulation/transport model to yield the geographic and seasonal variations of CO_2 in the atmosphere. The resemblance between the modeled and observed CO_2 distributions at the sparse remote observational network at the surface provides some but not very strong confidence in the surface fluxes, because of possible compensating errors in the model representation in the vertical and over continental regions.

Understanding the contemporary carbon budget requires information about how these fluxes have changed in response to the fluctuating climate and increasing atmospheric CO_2. Direct long-term measurements of the time-varying fluxes are lacking, and so time-varying atmosphere-land and atmosphere-ocean fluxes derived from the biogeochemcial models cannot be adequately validated. We have therefore relied on "inverse" modeling, with the aid of 3D atmospheric circulation/transport models, to infer the "perturbation" sources and sinks from the tracer CO_2 in the atmosphere. Because of the (incomplete) mixing by the atmospheric circulation, tracer CO_2 contains information about sources/sinks over broad regions only. Ancillary data are required to aid the interpretation. Different investigators have used different regional and temporal resolution as well as different ancillary data in their inversion studies, thus resulting in a range of scenarios about carbon sources and sinks. In the decade of inverse modeling of the contemporary carbon cycle, the existence of a land sink in the northern middle-latitudes is a robust conclusion that follows from the north-south gradient (~3–4 ppmv) in atmospheric CO_2 (Tans *et al*., 1990; Enting *et al*., 1995, Bousquet *et al*., 1996, 2000; Fan *et al*., 1998; Rayner *et al*., 1999). The partitioning of the land sink between North America and Eurasia is highly controversial (Fan *et al*., 1998; Rayner *et al*., 1999) and shown to be not robust because of the smallness of the east-west gradient of CO_2 between N-Atlantic and N-Pacific (<0.5 ppmv)

and the lack of continental data to constrain the result (Bousquet *et al.*, 1999a, b). CO_2 from tropical sources/sinks are not constrained by inversions, as they impact little north-south gradient, and their dilution by convection through the depth of the troposphere leave little local surface signature as well. Tropical sources/sinks are thus treated as residuals in the global budget, after the middle-high latitudes sources/sinks have been inferred from the tracer gradients.

Interannual variations in atmospheric CO_2 are large (~1–3 ppmv) compared to the east-west gradient, and so inversion for the interannual variations in sources and sinks are relative robust (Rayner *et al.*, 1999; Bousquet *et al.*, 2000). Fung (2001) found that the year-to-year variations of the northern and southern carbon sinks are related to the southern oscillation index (SOI), but with different phasing. The northern sink is in phase with the SOI, and is stronger during El-Nino's. The southern sink (mainly the ocean) lags the SOI by a year, and is weaker during El-Nino's. Randerson *et al.* (1999) found that both spring and autumn anomalies of atmosphere-land CO_2 fluxes at high latitudes are correlated with regional temperature anomalies. Again, the sign of the correlation coefficient varies with seasons: warming enhances both spring time photosynthesis and autumn decomposition. That these results are consistent with what is generally known about the sensitivity of CO_2 exchange processes to climate perturbations provide some confidence in the inverse modeling results in general. More importantly, these results yield quantitative relationships between carbon fluxes and climate fluctuations on a regional scale-relationships that may not be generalizable from local flux measurements. These relationships will be useful for evaluating models in which the carbon cycle is interactive with the climate.

MULTI-ELEMENT INTERACTIONS IN THE CARBON CYCLE

The fluxes of CO_2 among the atmosphere-land-ocean transfer also the isotopes of carbon and oxygen, and the exchange processes such as combustion, photosynthesis and respiration also alter the abundance oxygen in the atmosphere and oceans. Thus, the distributions of these other constituents provide additional information about the carbon cycle (e.g. Fracey *et al.*, 1995).

We illustrate the use of carbon-13 in the diagnosis of the contemporary carbon cycle in Fig. 1, which plots the annual C-13 flux against the annual C-12 flux. The slopes of the lines are given by the relatively well-known fractionation coefficients associated with each carbon flux. In the C-12 analysis the global carbon sink is calculated as the difference between the annual sources and annual change in atmospheric inventory. With the additional dimension given by C-13, the analysis can be extended to separate between land and ocean sinks, as these exchange processes have different degrees of discrimination against C-13. There are two complicating factors in the use of C-13 to constrain the contemporary carbon budget. The first relates to the isotopic disequilibria between incoming and outgoing fluxes because of the decreasing trend in atmospheric $\delta^{13}C$. The magnitudes of the disequilibria are related to the residence times of carbon in the terrestrial and oceanic reservoirs. The atmospheric C-13 signatures associated with the gross fluxes are thus large compared to the net fluxes (e.g. Fung *et al.*,

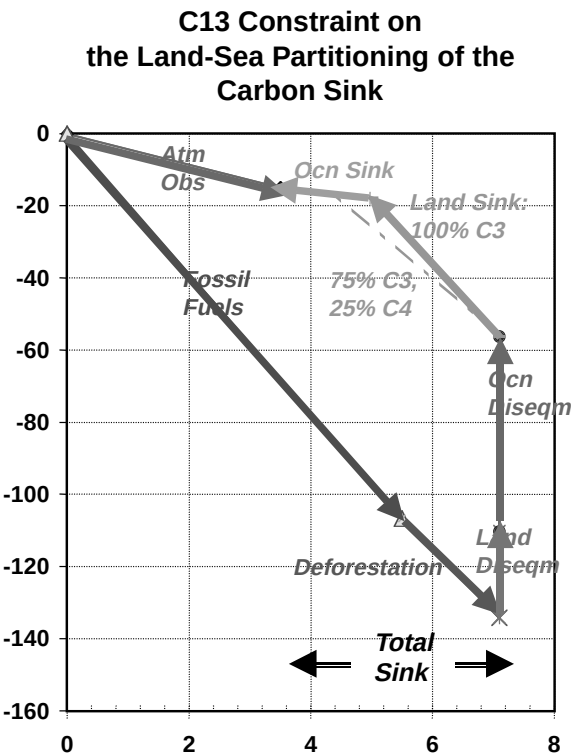

Fig. 1. Schematic diagram showing the C-13 constraint on the contemporary carbon budget.

1997) and remain a major uncertainty in the C-13 analysis. Resolution of the disequilibria will also provide confidence in estimates of the capacity of the land to sea to store anthropogenic carbon (cf. Eqs. (2) and (3)). The second complicating factor is the existence of C_4 vegetation (e.g. grasses, corn), whose photosynthetic discrimination resembles that associated with the oceans. Research is ongoing to quantify the seasonal and geographic variation in the areal extent of C_4 vegetation and their role in the carbon cycle (Still, 2000).

A goal for integrated carbon-climate models is to understand how interactions between carbon and climate modify the rates of CO_2 growth and climate change. An underlying expectation is that carbon sinks today may not be carbon sinks in the future. This is because turnover times of carbon in the different carbon reservoirs are sensitive to climate, as illustrated by the interannual variations of CO_2. Also, biogeochemical processes that do not play a central role in the carbon cycle now may dominate in a different climate. An example of such a process is iron limitation of marine productivity.

Regions of the oceans have excess macronutrients (NO_3 and PO_4) despite the abundance of sunlight. It is hypothesized that productivity is limited by

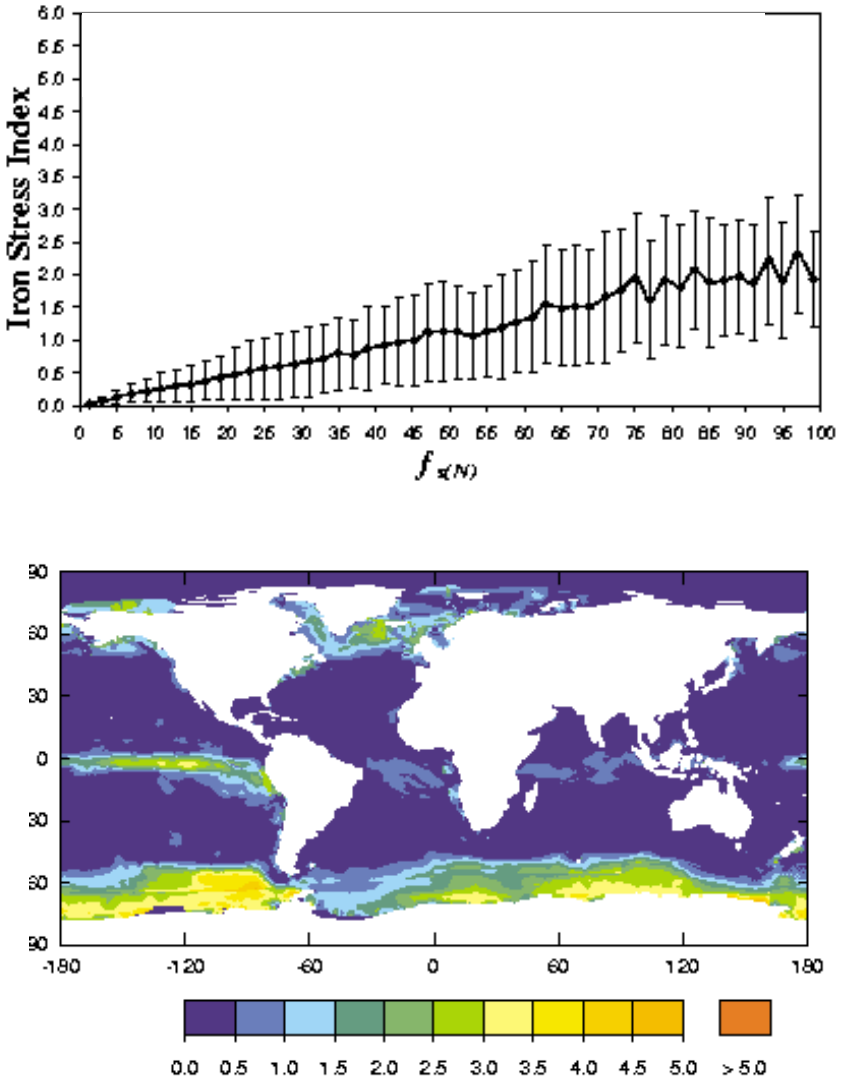

Fig. 2. Upper Panel: Iron stress index (see text for definition) as a function of $f_s(N)$, the ratio of N-utilization to N-supplied by upwelling. High values of $f_s(N)$ indicate low N-use efficiency. Lower panel: Geographic distribution of the iron stress index (from Fung *et al.*, 2000).

micronutrients such as iron. Aeolian deposition is a principal source of "external" iron. Dust lofted from arid regions are transported long distances and deposited to the oceans by rain or turbulent mixing (wet and dry deposition). Some of the iron in the dust is available for marine productivity. Enhancement of the productivity would likely depress pCO_2 of the surface water and lead to additional

uptake of carbon by the sea. This aeolian iron-marine productivity connection thus links the atmospheric CO_2 growth rate to terrestrial hydrology and marine biogeochemistry.

We have analyzed the iron stress in the ocean by combining estimates dust deposition from an atmospheric model, nutrient upwelling from an ocean model, and marine productivity from satellite observations to estimate the supply and demand of iron in the upper ocean (Fung *et al*., 2000). We defined an iron stress index as the ratio of Fe:N in marine phytoplankton relative to the ratio of Fe:N in the aeolian and upwelling supplies, and found that the iron stress index increases with decreasing N-use efficiency. In other words, marine productivity is greatest not only where there is abundant supply of NO_3, but also where there is efficient recycling of both N and Fe (Fig. 2).

FUTURE CHALLENGES OF MODELING GLOBAL ENVIRONMENTAL CHANGE

Projections of future climate change must be coupled with projections of future levels of CO_2. Making radiatively CO_2 a prognostic variable in global climate models (e.g. Cox *et al*., 2000) and including biogeochemcial interactions among carbon-nitrogen-iron and other elements are obvious next steps for which the present state of understanding may be adequate for a first undertaking. Technical issues such as coupling shock need to be addressed, but do not appear insurmountable.

As the global environment proceeds to a climate space for which the Earth has no prior experience, or for which we do not have adequate fossil records, it is difficult to validate the models and articulate our confidence in model projections of the future. An example is the fate of the tropical rainforest in the warmer, perhaps drier, but higher CO_2 world. The flourishing or decimation of the forest would have immense but opposite consequences for atmospheric CO_2 and climate. The challenge for modeling global environmental change is thus the lack of data. The data needs fall in two areas. The first involves observations of the present environment, so that models can be evaluated and improved. The second involves manipulative experiments so that environmental interactions in the future climate space can be parameterized with confidence in the models.

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